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MINERALIZATION AND HOST ROCK ALTERATION IN RELATION TO CLAY MINERAL DIAGENESIS AND EVOLUTION OF THE MIDDLE-PROTEROZOIC, ATHABASCA BASIN NORTHERN SASKATCHEWAN CANADA

by

J. Hoeve
D. Quirt

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ABSTRACT

We have studied the distribution of clay minerals in Middle-Proterozoic red beds of the Athabasca basin and in alteration haloes associated with unconformity-type uranium deposits by means of automated, quantitative XRD techniques. A coherent clay mineral stratigraphy is recognized which correlates with lithostratigraphic subdivisions and forms the basis for our interpretation of basinal development and diagenetic history. The application of clay mineral criteria to monitor diagenetic progress reveals intimate relationships between basin evolution, red bed diagenesis and the formation of the uranium deposits. Discrete stages of mineralization and remobilization are directly tied to stages of basin development and of prograde and retrograde diagenesis. We argue that hydrothermal mineralization at the unconformity is coupled to prograde diagenesis by free convection within the overlying sandstone aquifer.

The Athabasca basin constitutes a tectonically-active, intra-cratonic, sedimentary basin whose formation at 1450 ± 50 Ma was preceded by an episode of mild crustal uplift at approximately 1550-1500 Ma. Basin formation was initiated by an early tectonic rifting phase which induced block faulting related to pre-existing major Hudsonian faults, resulting in the formation of three sub-basins. Early stratigraphic development was controlled by the sub-basins as they were filled with fluviatile detritus derived from a kaolinite-rich provenance area underlain by a lateritic weathering profile. Towards the end of this period, when the sub-basins coalesced to form the larger Athabasca basin, the stratigraphic thickness was in the order of 1000 m. The ensuing non-tectonic, thermal relaxation phase was marked by a general downwarp and broadening

of the basin resulting in a basin-wide stratigraphic development and in overstepping of the earlier basin margins. The stratigraphic record of this phase comprises a cyclical shallow-marine to a fluviatile-terrestrial depositional pattern. Depth of burial is unknown but must have exceeded the presently preserved stratigraphic thickness of approximately 2 km.

Initiation of diabase intrusive activity at 1350 Ma marked the onset of a phase of renewed crustal instability and tectonic reactivation, lasting approximately 350 Ma from 1350 to 1000 Ma. During this period of active subsidence the basin reached its deepest level of burial and was subjected to peak diagenetic conditions. Total sediment thickness as inferred by previous workers from fluid inclusion studies was in the order of 4-5 km, which is consistent with clay mineral characteristics suggesting diagenetic conditions of deep burial and temperatures around 150-200°C.

Two prograde diagenetic trends, reflecting differences in original detrital mineralogy and sedimentary environment, are recognized in the clay mineral assemblages of the Athabasca Group rocks. A continental trend, expressed in formations of a fluviatile-terrestrial character, started with detrital kaolinite which, during burial, was variably converted into illite, depending on lithologically controlled potassium availability. A marine trend in intercalated shallow-marine silt- and shale-rich formations started with an association of smectite, mixed-layer minerals, glauconite, collophane and felsic tuff fragments, which during burial was transformed into a diagenetic, apatite-bearing, illite-chlorite assemblage.

Retrograde diagenesis, expressed by intensive fault-controlled kaolinitization, is attributed to a late period of uplift and erosion when surface waters regained access to the deep aquifer, following the Permian re-emergence of the Canadian Shield above sea level.

Primary ore formation took place during an advanced stage of basin development, coincident with the phase of tectonic reactivation starting at approximately 1350 Ma, which is 100-150 Ma after deposition of the Athabasca Group, in response to fluid interaction between the sandstone cover and the underlying basement, under conditions of deepest burial and highest grade diagenesis. Two main stages of mineralization are recognized which we infer to correlate with pulses of diabase magmatic activity accompanying episodic tectonic reactivation and active deepening of the basin. During the 1350-1250 Ma interval, diagenetic-hydrothermal systems of cover-basement interaction were established throughout the Athabasca basin, initiating the first and main stage of mineralization. This stage is characterized by the formation of high-grade, arsenide-facies ore, clay mineral alteration (illitization; chloritization), extensive quartz dissolution (carbonate when hosted by dolomitic basement rocks), graphite destruction, hydrocarbon formation and secondary hematitization. Basin-wide reactivation of the hydrothermal systems during the 1100-1050 Ma interval gave rise to the second stage of mineralization marked by sulphide facies ore and extensive superimposed bleaching. A clustering of U/Pb ages around 900-700 Ma may suggest a possible third stage of mineralization, which, however, has not been paragenetically defined and remains to be confirmed. Much younger remobilization at 300-250 Ma is associated with intensive

kaolinitization which is a manifestation of retrograde diagenesis related to Permian uplift and unroofing of the Athabasca basin.

A qualitative metallogenetic model proposes that diagenetic-hydrothermal systems of sandstone-basement interaction were driven by free convection within the basal Athabasca aquifer. Temporal heat flow variations associated with episodes of tectonic reactivation and mafic magmatism regulated free convection within the aquifer, thereby operating a switching mechanism for episodic diagenetic-hydrothermal activity on a basin-wide scale at intervals in the order of 100-200 Ma.

The distribution of convective cells and the location of the diagenetic-hydrothermal sub-systems of sandstone-basement interaction are inferred to have been strongly influenced by the distribution of graphitic basement rocks sub-cropping at the basin floor. The elongated linear shape of unconformity-type deposits in the Athabasca basin suggests that convective cells developed into longitudinal rolls that flanked the traces of graphite-bearing zones at the unconformity. Graphite exerted a dual control on the location of the diagenetic-hydrothermal flow systems: (1) a thermal control by virtue of its conductive properties, leading to the creation of a thermal anomaly at the unconformity which focussed convection over graphitic basement rocks, and (2) a mechanical control by virtue of its lubricant properties so that contemporaneous faulting, necessary to create a plumbing systems for sandstone-basement fluid interaction, was also focussed around graphitic basement rocks. In addition, graphite exerted a chemical control on the mineralization process by reacting with oxidized Athabasca pore fluids that penetrated below the unconformity, and converting these into reduced fluids containing hydrocarbons.

Ore formation resulted from dynamic interaction between the reduced fluids re-emerging above the unconformity and oxidized, metal-bearing pore fluids of the basal Athabasca aquifer and was, consequently, subjected to hydrodynamic control. The role of free convection, other than to drive the diagenetic-hydrothermal sub-systems, was to flush ore metals, released to the pore fluids in the course of red-bed diagenesis, from the basal Athabasca aquifer and transport these to sites of deposition at the unconformity. Significant ore, however, was precipitated only where a stationary redox front was maintained, as now outlined by a shell of secondary hematitization around pods of high grade ore. Barren clay alteration haloes resulted where no stationary redox boundary was established. The significance of convection is that it provides for highly focussed flow, so that under suitable hydrodynamic conditions, large, high-grade deposits could develop over time in diagenetic-hydrothermal systems characterized by a continuous replenishment of ore constituents and a continuous production of hydrocarbon reductants. The significance of a stationary redox front is that it provides for focussed precipitation.

In practical terms, the identification of criteria diagnostic of host rock alteration is very important because unconformity-type deposits present very small targets for exploration. Experience in the Athabasca basin has shown that host rock alteration criteria such as clay alteration, bleaching and quartz dissolution are very helpful (1) in evaluating the ore potential of a graphitic, conductive zone; (2) in identifying targets for diamond drilling; (3) as ore guides in evaluation drilling of known deposits. In the last application, a third

hydrothermal centre, as yet untested for ore, was identified at the Midwest deposit.

In a broader perspective we speculate that the interplay between burial diagenesis, convective flow and cover-basement interaction may be of greater significance to the formation of a variety of unconformity-associated and sediment-hosted ore deposits than is currently appreciated, because the Athabasca basin is unlikely to be unique. We consider that the geneses of unconformity- and sandstone-type uranium deposits are closely related in that these deposits may represent end products and transient, intermediate products, respectively, of a process of basin evolution and diagenesis that has gone to completion in the Athabasca basin. Analogous relationships may exist (1) between red-bed copper deposits and Proterozoic, giant stratiform copper deposits such as those of White Pine, Michigan, USA, and the African Copper Belt; (2) between certain types of sandstone lead deposits. The main difference in geological setting between the Proterozoic, giant stratiform copper deposits and unconformity-type uranium deposits may be that the redox front, in the case of the former, was located not at the underlying unconformity but at the top of the red bed aquifer at the base of overlying reduzate shales. Certain sandstone lead deposits, for example, those at the base of the Proterozoic Sibley Group in Ontario, Canada, and the much younger Maubach and Mechernich deposits in the northern Eifel, Germany, can be classified as unconformity-type deposits, because their formation appears to reflect cover-basement interaction long after deposition of the sediments. Possible other examples of unconformity-type deposits are two uranium deposits from undisclosed locations in the Soviet Union and the Co-As deposit of Bou Azer, Morocco.

From this view point, ore formation is an inherent aspect of basin evolution and diagenesis, which opens the prospect of developing a unified geological model for a range of sediment-hosted and unconformity associated metal deposits. Depending on whether the process of basin evolution goes to completion or is arrested at intermediate stages, a variety of deposits may be generated, ranging from those formed at low temperatures in a shallow environment shortly after deposition to those formed under conditions of deep burial and high-grade diagenesis during advanced stages of basin development. Thus, unconformity-type uranium deposits; stratiform uranium deposits such as in Gabon and Lodève; giant, stratiform copper deposits and unconformity-type lead deposits can be considered end members of a spectrum of metal deposits, whose unifying feature is an origin that is intimately tied to red-bed diagenesis.

The time-bound character of unconformity-type uranium deposits and stratiform copper deposits may suggest that crustal and tectonic conditions allowing the process of basin evolution, red bed diagenesis and ore formation to proceed to completion, have rarely been met in the earth's history. Unconformity-type uranium deposits are primarily known from the Middle-Proterozoic, whereas stratiform copper deposits have their main development in the Middle-Proterozoic and the Permian. These were periods when the land areas on earth were assembled in gigantic supercontinents, so that conditions were conducive to the formation of extensive intracratonic basins, to widespread fluviatile terrestrial sedimentation and to extensive red bed formation.

INTRODUCTION

Clay minerals are sensitive indicators of the low-temperature geological environment which may contain a record of the diagenetic and burial history of a sedimentary basin (Dapples, 1967; Muller, 1967; Dunoyer de Segonzac, 1969; 1970; Perry and Hower, 1970; Boles and Franks, 1979; April, 1981; Hurst and Irwin, 1982; Singer and Muller, 1983; Weaver and Brockstram 1984; Srodon and Eberl, 1984) or of host rock alteration associated with certain types of ore deposits (Heyl et al., 1964; Kerr and Jacobs, 1964; Iijima, 1974; Jambor and Delabio, 1978; McLeod and Stanton, 1984).

Pioneering studies on the diagenetic evolution of clay minerals were made by Burst (1959) and Weaver (1959, 1960), who by means of XRD techniques documented progressive clay mineral transformations with increasing depth of burial. Since then, the burial history of clay minerals has been traced for both shales and sandstones in various sedimentary basins and general patterns have been recognized (Velde and Hower, 1963; Dunoyer de Segonzac, 1970; Hower et al., 1976; Hower, 1981; Kish, 1983). It appears that smectites are converted into illite and chlorite via intermediate stages marked by random and regularly ordered smectite-illite and smectite-chlorite interlayering. Kaolinite is directly replaced by illite and chlorite, except in alkali- and Fe-Mg-poor aluminous sediments, where the mineral may persist until converted into dickite. The crystallinity of illite increases and its polytype changes from 1Md to 1M to 2M. Chlorites tend to change from Fe-rich to Mg-rich compositions and to develop Ib(β = 90°) and IIb(β = 97°) polytypes, although in Fe-Mg-poor, aluminous sandstones di,dioctahedral (donbassite) and di,trioctahedral (sudoite) varieties may form. Conco-

mitantly, unstable detrital silicates such as micas, feldspars and ferromagnesian minerals are progressively destroyed. As a result of these studies, clay minerals have become very useful monitors of diagenetic progress, particularly following calibration against other indicators of diagenesis and incipient metamorphism, including vitrinite reflectance and mineral assemblages of the zeolite- and prehnite-pumpellyite facies (Frey and Niggli, 1971; Kübler et al., 1979; Kish, 1980,a, b; Frey et al., 1980). Clay mineral criteria of diagenesis have since been applied to a variety of geological problems ranging from the interpretation of basinal history (April, 1981; Hamoumi et al., 1981; Hurst and Irwin, 1982) to tectonic reconstructions (Maxwell and Hower, 1967; Brasier et al., 1979; Aprahamian and Pairis, 1981; Kish, 1981), the detection of early, low-grade effects of regional metamorphism and orogenesis (Islam et al., 1982; Duba and Williams-Jones, 1983a; Thompson and Frey, 1984), the recognition of low-grade thermal effects of contact metamorphism and hydrothermal mineralization (Duba and Williams-Jones, 1983b) and of retrograde metamorphism (Lippmann and Rothfuss, 1980).

In petroleum geology, the use of clay minerals is widespread, both in exploration and production (Longstaffe, 1981) as well as in more fundamental studies that trace the relationships of organic matter maturation and hydrocarbon generation and migration to basin evolution and diagenesis (Burst, 1969; Foscolos et al., 1976; Héroux et al., 1981). This is in striking contrast to the situation in ore geology where the use of clay minerals has largely been restricted to the sporadic delineation of alteration haloes (Heyl et al., 1964; Kerr and Jacobs, 1964; Iijima, 1974; Blount et al., 1984; McLeod and Stanton,

1984). We are not aware of any studies that have investigated the interrelations of basinal history, diagenesis and mineralization, even though diagenesis is often invoked as a factor in the formation of sediment-hosted ore deposits (Wolf, 1976; Gustafson and Williams, 1981).

The present report will discuss the application of clay mineral criteria to problems of uranium metallogenesis in the Middle-Proterozoic Athabasca Basin in northern Saskatchewan, where previous work had implicated diagenesis of Athabasca Group red beds in the origin of unconformity-type uranium deposits (Pagel, 1975; Hoeve and Sibbald, 1976; 1978; Hoeve et al., 1980).

Under a four year program, studies of clay mineral host rock alteration around individual deposits were complemented by regional studies of clay mineral distribution and diagenetic patterns in the Athabasca Group rocks. Objectives were twofold: firstly, to gain an understanding of the interrelationships between diagenesis, host rock alteration and mineralization and, secondly, to identify diagnostic ore guide criteria that could be applied to drill hole evaluation in exploration programs.

To date, approximately thirty thousand samples have been analyzed for clay minerals by means of automated, quantitative XRD techniques at the Saskatchewan Research Council. Initial samples were collected and analyzed as part of our ongoing research program, but the majority were subsequently submitted by industry, after the significance of clay mineral analysis to exploration had been demonstrated in a number of pilot studies (Hoeve, 1981; 1982 a,b,c; 1984). The analytical capacity of SRC's clay mineral laboratory is in the order of 2000 samples per

month, thanks to the fact that the laboratory to a large extent is dedicated to Athabasca samples. It is on account of these circumstances that a study of the present scope and magnitude became practical to undertake.

SAMPLING, SAMPLE PREPARATION AND ANALYTICAL PROCEDURES

For the regional study of clay mineral distributions and diagenetic patterns in the Athabasca group, samples were taken at 5 and 10 metre intervals from available stratigraphic drill cores located throughout the Athabasca basin (Figures 1,2). Several of these cores had been previously examined by Ramaekers (1979; 1980; 1981), thus allowing direct comparison of clay mineral data with lithostratigraphy. For the host rock alteration studies, selected drill cores forming fences across the strike of individual deposits were similarly sampled at 5 and 10 metre intervals. Twelve fences were studied for the Midwest deposit, and two for each of the Key Lake, Collins Bay B and McClean deposits. The Midwest and Collins Bay B cores were logged by us in detail, whereas we had to rely on company logs for the other deposits.

Sample preparation includes crushing and grinding followed by the extraction of the $\leq 2\mu m$ size fraction by means of centrifuging techniques. The procedure is detailed in Jambor and Delabio (1978) and is similar to the one employed by the Geological Survey of Canada. The extracted clay mineral slurry is transferred to glass slides by the smear-on technique, which results in oriented mounts suitable for (semi-) quantitative analysis (Gibbs, 1965).

The XRD analysis is based upon the relationship

$$\frac{I_i}{I_j} = K \frac{x_i}{x_j}$$ (Brindley, 1961),

where K represents a constant which includes factors related to the nature and density of the components i and j and the geometry of the XRD equipment. In the analysis, intensity measurements of selected peaks are made and intensity ratios $\frac{I_i}{I_j}$ calibrated by means of the experi-

mentally determined factor K, to give the weight ratios $\frac{x_i}{x_j}$ between the various clay minerals in the sample. For convenience, these are then converted to relative weight percentages.

Monomineralic reference materials were selected during an early stage of the program from the samples that were submitted for analysis, including kaolinite, illite and di,trioctahedral chlorite (sudoite). As the program continued, the reference collection was expanded with trioctahedral iron- and magnesium-rich chlorites, dickite septechlorite (berthierine), and illite/smectite containing ~ 10% expandable layers. Standard mixtures were made up from these reference materials, so that the characteristics ("crystallinity", composition, polytype) of the standards approach as closely as possible those of the unknown samples to be analyzed.

The following measurements/determinations were made:

(1) Identification of minerals present.

(2) Kübler index of illite "crystallinity" (Kübler, 1968).

(3) The intensity ratio I(002)/I(001) of illite.

(4) Selected samples were tested for the presence of expandable phases by means of ethylene glycol treatment. Srodon's (1984) criteria were applied to estimate the percentage of smectite layers in illite.

(5) Illite polytype determinations on unoriented mounts of selected samples, using XRD-data of Bailey (1980) and Carroll (1970) and criteria reported by Srodon and Eberl (1984).

(6) Identification of chlorite type: di,trioctahedral (sudoite) or trioctahedral (necessary for correction of kaolinite/

chlorite peak overlaps). In order to minimize effects of peak intensity variations for trioctahedral chlorites of intermediate composition, the values used in the calculations were averaged over four peak areas (001; 002; 003; 004).

The Kübler index of illite "crystallinity" is measured as the width in millimetres at half height of the (001) reflection (note inverse relationship: the index decreases as the crystallinity increases). The value of this empirical unit also depends on the type and setting of the XRD equipment used. This means that measuring conditions must be standardized and, furthermore, that interlaboratory comparisons are only possible after interlaboratory calibration of results. The SRC measurements are calibrated against standard samples supplied by Prof. Kübler's laboratory, Neuchâtel, Switzerland.

To evaluate the effects of grinding on the value of the Kübler index, a comparison was made between batches of samples that were crushed and ground and those that were gently disintegrated by means of ultrasonic treatment. As no significant differences were noted, grinding was adopted as the most effective technique of rock disaggregation.

GENERAL GEOLOGY

The Precambrian Shield of northern Saskatchewan (Figure 1) is made up of Archean and lower-Proterozoic rocks which have been variably affected by Hudsonian orogenesis (≈1800 Ma) and form a metamorphic basement to unconformably overlying, unmetamorphosed red beds of the Middle-Proterozoic Martin and Athabasca Groups (terminology after Ramaekers, 1979, 1980, 1981; Langford, 1981). The basement, as well as the Martin and Athabasca Groups, are intruded by pre-Grenvillian diabase dikes, which generally follow NW-SE directions. At Beaverlodge, some of the dikes dated at 1835±50 Ma (Wanless et al., 1966), may have acted as feeders for volcanic flows intercalated with the Martin red beds. In the Athabasca basin, diabase dikes were intruded during the interval of approximately 1350-1000 Ma (Bell, 1981; Armstrong and Ramaekers, 1985). The unconformity is underlain by a partially preserved palaeoweathering profile of lateritic character, which is in the order of 10-20 m thick and displays a conspicuous colour zonation (Hoeve, 1977; Hoeve and Sibbald, 1978; MacDonald, 1980). The upper part is oxidized and coloured a pervasive red or purple by hematite pigment, whereas the lower part is reduced and predominantly greenish. The transition between these two zones defines a basin-wide redox boundary between oxidized Athabasca red-bed sandstones above the unconformity and reduced basement rocks below.

The metamorphic basement is transected by several large NE-SW trending fault zones which separate major crustal units (Figure 1) and which remained active for a long time after the Hudsonian orogenesis. These played an important role in the initiation and subsequent development of the Athabasca basin (Ramaekers, 1979; 1980; 1981).

THE ATHABASCA GROUP

The Athabasca basin, underlying an area of approximately 100,000 km^2, is filled with red-bed sandstones, conglomerates and minor shales and dolomites of the Middle-Proterozoic Athabasca Group. The area was first mapped by Fahrig (1961) who considered the red beds to consist of ultra-mature, poly-cyclic, orthoquartzites of a fluviatile nature. Ramaekers (1979; 1980; 1981) distinguished several lithostratigraphic formations (Figure 2) and interpreted the clastics to represent first-cycle, fluviatile and marine sediments which owe their apparent mineralogical maturity to peculiarities of the vegetationless Proterozoic weathering and depositional environments, and to a diagenetic history characterized by intrastratal alteration of detrital framework silicates to clay minerals.

On the basis of sedimentological and stratigraphic evidence Ramaekers (op. cit) proposes that the Athabasca Group was deposited in a tectonically active sedimentary basin located in a continental margin structural setting, whose development was initiated by the formation of three NE-SW trending sub-basins (Figure 3) controlled by the major Hudsonian faults in the metamorphic basement. These basins were quickly filled in with detritus and coalesced to form the larger Athabasca Basin, whereas the faults remained active during sedimentation and controlled to some extent later basin development. Diabase dikes intruded the clastic sequence during the interval 1350-1000 Ma, which is interpreted as a period of subsidence and deep burial of the sediments. The general NW-SE trend of the dikes corresponds to tensional directions associated with left-lateral movements along the Hudsonian faults (Ramaekers, 1980; 1981; Ramaekers and Hartling, 1978).

Table 1. Summary of published Rb/Sr, K/Ar and Ar^{40}/Ar^{39} dates for unaltered Athabasca Group rocks and for alteration clays.

A. Unaltered Rocks	Rb/Sr	K/Ar	Ar/Ar	Source
Shale Wolverine Point	1340 ± 50 Ma			Ramaekers and Dunn, 1976
Tuff Wolverine Point	1428 ± 30			Ramaekers, 1980
Tuff Wolverine Point	1450 ± 30			Bell and Blenkinsop, 1980
Tuff Wolverine Point	1430 ± 30			Armstrong and Ramaekers, 1985
Illite in Douglas and Lazenby Lake sandstone		1222 ± 52		Clauer et al., 1985
Illite in Manitou Falls sandstone		1350 ± 33		Wilson et al., in press
Illite in Manitou Falls sandstone		1084 ± 37		Wilson et al., in press
B. Altered rocks				
Illite Key Lake		1257 ± 24		Wilson et al., in press
Illite Key Lake		1187 ± 23		Wilson et al., in press
Illite Key Lake		1189 ± 38		Wilson et al., in press
Illite Key Lake		1131 ± 24		Wilson et al., in press
Illite Key Lake		1154 ± 22		Wilson et al., in press
Illite Midwest		1039 ± 22		Wilson et al., in press
Sericite Midwest	1337 ± 11			Worden et al., 1986
Sericite Midwest	1222 ± 28			Worden et al., 1986
Sericite Midwest	1249 ± 48			Worden et al., 1986
Illite Cluff		1293 ± 36		Clauer et al., 1985
Illite Collins Bay		1266 ± 39		Stevens, et al., 1982
Illite Collins Bay		1296 ± 34		Stevens, et al., 1982
Illite McLean			1320 - 1170	Hall et al., in press

The Athabasca Group sediments themselves have been variably dated (table 1) by means of Rb/Sr methods at 1340±50, 1428±30, 1513±24 and 1450±30 Ma, with the last age probably giving the best estimate (Ramaekers and Dunn, 1977; Ramaekers, 1979, 1980; Bell, 1981; Armstrong and Ramaekers, 1985). Diagenetic ages, as determined by K/Ar-dating of authigenic illites, are much younger, in the range of ~1350-1050 Ma (Clauer et al., 1985; Wilson et al., in press).

The preserved part of the stratigraphic column is at least 2 km thick; however, fluid inclusion evidence (Pagel, 1975a, b) suggests an original thickness of the sedimentary cover of approximately 5 km, which is consistent with the results of clay mineral studies indicating deep burial and high-grade diagenesis (Hoeve et al., 1981).

The basal part of the stratigraphic column (table 2), as interpreted by Ramaekers (1979, 1980, 1981), consists of two clastic wedges which in places are underlain by debris flow deposits of local derivation (Ramaekers, 1981; Pacquet, 1985). These wedges, comprising the Manitou Falls Formation (MF) in the Cree and Mirror basins and the Fair Point Formation (FP) in the Jackfish basin (Figures 2,3) were laid down upon a deeply weathered lateritic surface. The Manitou Falls Formation, subdivided into four members (MFa, MFb, MFc and MFd) consists of coarse, fluviatile conglomerates and pebbly sandstones (MFa + b) overlain by sandy, braided-stream and wadi deposits (MFc + d), which were deposited from the east. The latter commonly contain clay and silt intraclasts that are deformed by compaction. The lower members (MFa + b) are restricted to the Cree basin, whereas the upper members (MFc + d) extend westwards into the Mirror and Jackfish basins (Figures 2,3). The Fair Point Formation (FP), stratigraphically equivalent to the lower Manitou

Table 2. Generalized stratigraphy of the Athabasca Group: (1) Ramaekers, 1980; (2) This volume.

Formation	Lithology	Thickness (m)	(1) depositional environment	(2)
Carswell	stromatolitic, oolitic dolomite	>300	marine	marine
Douglas	Interbedded carbonaceous siltstone and shale fine-grianed sandstone; mud cracks	≈500?	marine	marine
Tuma Lake	Cross-bedded, pebbly sandstone	100	marine	marine, transgressive (?)
Otherside	Fine-grained, well-sorted, cross-bedded quartz sandstone	350	marine	fluviatile
Locker Lake	Massively bedded, pebbly sandstone	<200	marine	fluviatile
Wolverine Point (upper member)	Siltstone and shale interbedded with medium-grained, clay-rich sandstone; phosphorite; glauconite; vitric tuff.	<300	marine	marine
Wolverine Point (lower member)	Medium-grained, cross-bedded sandstone, minor interbedded siltstone	<200	marine	marine, transgressive
Lazenby Lake	Cross-bedded, pebbly sandstone, lacking clay intraclasts	>600	marine	fluviatile
Manitou Falls	Cross-bedded, medium-grained, pebbly sandstone. Lower members interbedded with conglomerate. Upper members consist of pebble free, finer-grained sandstone, carrying abundant clay intraclasts	<1400	fluviatile } stratigraphically equivalent formations	fluviatile
Fair Point	Coarse-grained, clay-rich, pebbly and conglomeratic sandstone	300	marine } stratigraphically equivalent formations	fluviatile
	Unconformity			

Falls Formation (MFa + b), is interpreted by Ramaekers as a marine sandstone deposited in a high-energy, near-shore environment. This is contested by Wilson (1985), who does not find any evidence of a marine influence and advocates a continental, fluviatile depositional environment. The formation consists of clay-rich sandstones in which clay granules, interpreted as altered feldspars, are common, particularly near the base.

These two basal clastic wedges are overlain by a succession which Ramaekers (op. cit) interprets as near-shore to shallow-shelf, marine sandstones, comprising several transgressive-regressive cycles related to repeated uplift in the source areas to the east: Lazenby Lake (LzL), Wolverine Point (WP), Locker Lake (LL), Otherside (OF) and Tuma Lake (TL) Formations. In Ramaeker's stratigraphic reconstruction the Lazenby Lake Formation (LzL) is restricted to the western and central sub-basins and consists of fine- to medium grained, locally pebbly sandstones, which were transported from the south. The Wolverine Point Formation, subdivided into lower and upper members (WPa and WPb, resp.) is present throughout the basin, conformably overlying the Lazenby Lake and upper Manitou Falls Formations. The lower member (WPa) consists of sandstone, commonly containing clay/silt intraclasts, and minor intercalated siltstone. The upper member (WPb) is made up of clay-rich sandstones, siltstones and mudstones, containing interbedded, phosphatic layers. Clay/silt intraclasts are abundant, as are phosphatic pellets and fragments of felsic tuff. Intercalated silt layers are at least in part of tuffaceous origin, because glass shard textures can be recognized in many of them. These sediments were deposited in an inner to open shelf environment. The overlying Locker Lake, Otherside and Tuma Lake Forma-

tions consist of relatively pure sandstones which Ramaekers considers to have been deposited in an environment ranging from upper shore-face to inner shelf. Wilson (1985) disagrees and favours a continental, fluviatile environment.

The uppermost part of the stratigraphic column, consisting of shales and stromatolitic dolomites of the Douglas (DG) and Carswell (CW) Formations, are only preserved within the Carswell Structure, which represents a Silurian meteorite impact crater (Figure 1,2,3; Pagel, 1975; Ey, 1984). Contacts of the Douglas with the underlying Tuma Lake and overlying Carswell Formations are gradational.

The overall stratigraphic succession, from fluviatile to near-shore marine sandstones, to marine stromatolitic dolomites, is similar to that reported for the Thelon Formation in the Northwest Territories (Cecile, 1973; Donaldson, 1967) with which the Athabasca Group is often correlated (Young, 1984).

Petrography

In most of the Athabasca Group sandstones and conglomerates, quartz is the only detrital mineral of significance. Relict grains of altered biotite, altered lithic fragments and altered feldspar, though encountered throughout the sequence, are rare, except in the upper Wolverine Point Formation (WPb) where an immature detrital mineral assemblage is locally preserved by an early-diagenetic phosphate cement (F-apatite). Heavy mineral bands, characterized by Fe-Ti oxides, minor tourmaline and zircon are only abundant in the basal, conglomeratic portions of the lower Manitou Falls Formation (MFa + b) and to a lesser degree also in the Locker Lake Formation (LL). Fragments and, more rarely, layers of

felsic tuff up to a few metres thick (e.g. drill hole BM-1), first reported for the upper Wolverine Point Formation (WPb; Ramaekers, 1980), are now also recognized at the base of the sequence (Pacquet, 1985) as well as near the top in the shales of the Douglas Formation (DG; Lain , pers. comm.). Phosphatic minerals of the crandallite group are found in trace quantities throughout the entire Athabasca Group sequence, whereas diaspore is locally abundant within thin zones.

Detrital quartz consists primarily of monocrystalline grains, which are sub-rounded to well-rounded and poorly to well sorted. The packing density is variable as indicated by sutured, concavo-convex, point contacts and floating grains.

The interstitial clay content of the sandstones and conglomerates ranges between 5-25% in the basal portions of the lower Manitou Falls and Fair Point Formations to approximately 5-10% higher in the sequence with the exception of the upper Wolverine Point and Lazenby Lake Formations. The former is distinctly clay and silt rich, whereas the latter stands out by its low clay content (<2%).

Based upon chemical composition, plotted as log (SiO_2/Al_2O_3) versus log (Na_2O/K_2O), the majority of the sediments are classified as quartz arenites and sub-lithic arkoses, except for the sandstones of the upper Wolverine Point (WPb) and the lower Fair Point Formation which plot within the arkose and sub-arkose fields (Ramaekers, 1981; Quirt, 1985).

The constitution of the Athabasca Group rocks varies from incoherent to well-indurated, with authigenic quartz forming the most common cement in addition to interstitial clay and hematite. Subordinate carbonate cement (siderite) is characteristic of the Locker Lake formation but is rarely found in the other lithological units. An early-

diagenetic phosphate cement (F-apatite) is restricted to certain horizons within the upper Wolverine Point formation. Goyazite (Sr-hydrous-phosphate) is wide-spread, but present only in small quantities in the interstitial clay matrix. Diaspore is found at various stratigraphic levels within the Lazenby Lake (LzL) and Locker Lake (LL) Formations, forming a cement of poikiloblastic crystals enclosing detrital grains of quartz (plate 1A).

Authigenic quartz cement and interstitial clay exhibit antipathetic relationships, quartz overgrowths being best developed and most abundant in sandstones that contain little or no clay such as those of the Lazenby Lake Formation. In clay-rich rocks of the upper Wolverine Point Formation, quartz overgrowths are all but absent. Sandstones with extensive quartz cement are typified by floating grains and point contacts between detrital grains, indicating that the overgrowths formed early in the burial history, before significant compaction had taken place (plate 1B). Characteristically, many of the more intensively silicified rocks are marked by a uniform grain size, high roundness and sphericity and by a paucity of clay and heavy minerals, rendering a "ping pong ball" texture. Such features, which are particularly characteristic of the Lazenby Lake Formation, and to a lesser degree also of the lower Wolverine Point and upper Manitou Falls Formations, point to aeolian activity in the depositional environment (Glennie, 1972; Turner, 1980).

In the majority of the sandstones and conglomerates, clay occurs as a fine-grained, recrystallized, interstitial matrix, which is generally interspersed with specular hematite. In addition, coarser grained, partially deformed clay granules are present which were probably derived

from post-depositionally altered feldspar grains because relicts of original feldspar can occasionally be recognized. Such granules are common in the lower part of the Fair Point Formation, in the upper Wolverine Point Formation and in the Locker Lake Formation. Within the Manitou Falls Formation, they are rare and generally severely deformed by compaction, so that identification as original feldspar may be problematic. Radially textured grain coatings and pore linings of authigenic clay are only found within the upper Wolverine Point Formation (WPb), as are rounded clay pellets of illite and chlorite which may represent diagenetically modified glauconite grains.

The clay matrix of the sandstone consists of illite and chlorite in the upper Wolverine Point Formation, whereas in the other formations the clay is primarily composed of kaolinite and minor illite. In the latter, kaolinite forms a dense structureless groundmass, which may represent recrystallized original detrital clay. Kaolinite also occurs as coarse vermicular aggregates filling pore spaces and as rare pseudomorphous replacements of mica and feldspar. Illite is primarily present as fine-grained replacements of kaolinite.

Fragments of felsic tuff, common in silty intercalations of the Douglas (DG) and upper Wolverine Point (WPb) formations and also of the basal, locally derived debris flow deposits, are identified by diagnostic shard textures (plate 1C). Some of the tuff fragments are now composed of quartz and K-feldspar, or fine-grained quartz only, while most consist of clay (illite). Intraclasts and granules of dense fine-grained F-apatite are occasionally found in the sandstones of the Locker Lake Formation (LL), and are identical in appearance to phosphate concentrations in the upper Wolverine Point formation (WPb).

Hematite is ubiquitous throughout the sequence, imparting purple and dark red to brick red or pink colours to the rocks, depending on concentration and grain size. It is particularly abundant in the basal conglomeratic portion of the lower Manitou Falls (MFa + b) and Fair Point (FP) Formations, where it is present as coarse specularite in pore spaces, colouring the rock a dark to blackish purple. Liesegang rings are locally recognized in these purple rocks.

The Athabasca Group rocks are variably affected by superimposed chemical reduction of iron, manifested by zones having a bleached or a mottled appearance. Such zones which tend to be fracture and fault controlled, are very common in the lower Manitou Falls Formation, where exposures along the eastern edge of the basin show complicated but coherent patterns of bleaching and re-hematization. Purple, pervasively hematitized rocks are transected by broad, often irregular, bleached zones in the order of hundreds of metres wide, from which diagenetic specular hematite has been removed. Within the bleached zones re-hematization is manifested by large systems (tens of metres) of bright red liesegang rings made up of fine-grained hematite pigment, which cut across sedimentary bedding and also penetrate a short distance into the adjacent uniformly purple-coloured rocks. These sets of bright red rings are in turn affected by renewed bleaching, represented by haloes around cross-cutting joints and fractures. Recent re-oxidation is indicated by development of yellow-brown limonite rings, superimposed upon the fracture-controlled bleaching haloes.

In thin section, hematite occurs in several modes: (1) fine-grained coatings around detrital quartz grains, overgrown by oriented authigenic quartz rims. These coatings are absent at grain contacts,

thus attesting to their post-depositional, early diagenetic, pre-silicification origin (plate 1B); (2) coatings outlining growth zones within authigenic quartz rims; (3) dispersed, generally coarser-grained, specularite within the interstitial clay matrix.

Hematite-clay pseudomorphs after detrital mica, though found throughout the sequence, are rare in units other than the upper Wolverine Formation.

Detrital heavy mineral bands (black sands) are common in the basal portion of the Manitou Falls Formation (MFa + b) and less so in the Locker Lake Formation (LL). The opaque grains are composed of hematite, but bear evidence of having originated by oxidation of precursor Fe-Ti-oxides. Exsolution lamellae of rutile following rhombohedral directions indicate that the majority of the grains are pseudomorphous after ilmenite. Many pseudomorphs contain cores of goethite-rutile in optical continuity with the surrounding hematite-rutile (plate 1D). A minority of the grains is pseudomorphous after magnetite, as attested to by relict (111) directions. All but a few of the detrital Fe-Ti-oxide grains carry syntaxial overgrowths of specular hematite (plate 1D). Within the above-mentioned bleached zones, the hematite component of the pseudomorphs has been replaced by clay or siderite, giving rise to clay-rutile and siderite-rutile pseudomorphs after ilmenite (plate 1E).

Variably coloured detrital tourmaline grains, included in the heavy mineral bands, carry oriented overgrowths of a colourless tourmaline (dravite), which appears to be optically identical to the spherulitic, hydrothermal dravite found in alteration haloes around the uranium ore deposits. Whispy, tufts of radiating tourmaline needles are found in pore spaces, originating from notches on detrital quartz surfaces.

In interpreting these petrographic characteristics, which are summarized in table 3, there is no doubt that post-depositional alteration of detrital feldspar and mica to kaolinite and illite has taken place. Nonetheless, Ramaekers' (1980; 1981) generalization that this process has significantly contributed to the development of interstitial clay and hematite pigment, is questionable in view of the rarity of pseudomorphs after and relicts of detrital silicates in formations other than the upper Wolverine Point. Textural relationships suggest that illite probably formed by direct replacement of kaolinite.

The diagenetic history of iron appears complex, as is the case in many ancient continental red beds (Turner, 1980). If the goethite-rutile cores of the hematite-rutile pseudomorphs after ilmenite may be considered armoured relicts of an earlier phase in the oxide evolution, then the diagenetic history of the Fe-Ti-oxides can be reconstructed as follows. Initially, ilmenite grains were subjected to oxidation and hydrolysis, resulting in the formation of goethite-rutile pseudomorphs after the original mineral. This must have taken place under low-temperature conditions (Berner, 1969; Langmuir, 1971), possibly during lateritic weathering in the source area of the sediments prior to transportation and deposition. During burial and diagenesis, goethite underwent topotaxial replacement by hematite to form hematite-rutile pseudomorphs. Dispersed, colloidal limonite, which by analogy with modern lateritic terranes must also have been supplied by the provenance area (Walker, 1974; Riezebos, 1979), recrystallized to form the specular hematite pigment in the interstitial matrix as well as authigenic overgrowths around detrital hematite-rutile pseudomorphs (plate 1D). This interpretation is supported by the presence of liesegang rings in purple

Table 3. Idealized summary of paragenetic relationships in fluviatile sandstones of Athabasca Group.

Detrital Mineral	Prograde Diagenesis	
Quartz		oriented quartz overgrowths
Feldspar	Replaced by kaolinite	replaced by illite
Mica	Replaced by kaolinite and/or illite	replaced by illite
Limonite	hematite pigment	specular hematite in clay matrix
Ilmenite	goethite/rutile → hematite/rutile	
Magnetite	goethite → hematite	oriented overgrowths of specular hematite
Tourmaline	dravite overgrowths plus radiating tufts	
Kaolinite/?/Halloyzite?	recrystallization to dense matrix kaolinite	coarse "books" of authigenic kaolinite

coloured rocks, suggesting that the specular hematite was formed diagenetically from colloidal, limonitic precursors, which underwent dehydration and recrystallization into specularite during burial.

During a later phase of diagenesis, iron was subjected to repeated oxidation-reduction cycles, as recorded by the bleaching and re-hematization patterns described for the lower Manitou Falls Formation. The most recent oxidation phase is represented by liesegang rings of yellow-brown limonite.

Clay Mineral Distribution and Stratigraphy

For the regional study of clay mineral distribution and diagenetic patterns within the Athabasca Group, three sets of samples were analyzed: (1) sandstones and conglomerates; (2) intercalated clay and siltstone layers; (3) clay and silt intraclasts. These were collected from the same deep stratigraphic drill cores as had been used by Ramaekers (1979; 1980; 1981) in his sedimentological and stratigraphic studies (Section A-A', Figure 2), allowing a direct comparison between clay mineral stratigraphy and lithostratigraphic units. Additional drill core was sampled from the previously little known area to the south and southeast of the Carswell Structure and from the area near the northern rim of the basin (Sections B-B' and B'-B", Figure 2). The highest lithostratigraphic units, the Douglas and Carswell Formations, were sampled from a few drill cores located within the Carswell Structure (Figure 3; CAR 179; CAR 181). Supplementary information was obtained from drill cores in the Alberta part of the basin, which we had analyzed for the Alberta Research Council (Wilson, 1985).

Previous reports have demonstrated the use of clay minerals in stratigraphic interpretations of the Athabasca Group and discussed the

correlation between stratigraphic units inside and outside the Carswell Structure (Hoeve et al., 1981; 1985).

Sandstones and Conglomerates

The clay matrix of the Athabasca Group sandstones and conglomerates consists of varying proportions of kaolinite, illite and chlorite. Dickite is locally present as fracture fillings and associated impregnations. Accessory minerals include rutile, anatase, hematite, zircon, tourmaline, apatite, goyazite and locally diaspore.

The > 2 µm fraction consists primarily of kaolinite and no stratigraphic subdivisions can be made except for the Upper Wolverine Point Formation which stands out by an assemblage of illite and chlorite. By contrast the clay mineral logs for the ⩽2µm fraction of the analyzed drill holes (Appendix A) can be subdivided into depth intervals that correlate with distinct lithostratigraphic units (Hoeve et al., 1981). Thus, a general clay mineral signature for these units can be determined (Table 4). The Fair Point Formation in the western portion of the basin is characterized by a mineral assemblage of almost exclusively kaolinite (± illite). The Manitou Falls Formation in the eastern and central sub-basins contains kaolinite and illite in approximately equal proportions, with the lower members (MFa + b) being slightly kaolinite-dominated and the upper members (MFc + d) illite-dominated. The overlying lower Wolverine Point Formation (WPa) shows predominance of illite with intermediate kaolinite and generally accessory chlorite, whereas the upper Wolverine Point Formation (WPb) is distinguished by high proportions of both chlorite and illite, combined with an absence of kaolinite. The Locker Lake Formation (LL) is characterized by extreme illite-predominance, while the boundary against the upper Wolverine Point

Table 4: Clay Mineral Data for Athabasca Group Stratigraphic Subdivisions

		Sandstone Samples							
		LL + OF	WPb	WPa	LzL	MF c+d	MF a+b	FPb	FPa
Kaolinite	n	180	153	83	201	207	70	34	38
	$\bar{X}$ (%)	9.49	3.67	35.09	2.49	34.88	55.62	89.96	72.00
	95% conf. int. (%)	±2.623	±2.634	± 4.671	±1.047	±2.404	±4.239	±3.885	±5.947
	δ n-1 (%)	17.78	16.44	21.26	7.51	17.50	17.65	10.97	17.85
	RSD	187.47	447.38	60.60	301.14	50.17	31.74	12.19	24.80
	Range (%)	0.0 - 98.0	0.0 - 93.4	0.0 - 94.8	0.0 - 50.0	0.0 - 90.7	0.0 - 91.7	32.0 - 100.0	0.0 - 100.0
Chlorite	n	180	153	83	201	207	70	34	38
	$\bar{X}$ (%)	1.26	46.04	4.62	0.23	1.52	1.56	0.14	1.53
	95% conf. int. (%)	±0.747	±4.501	1.645	±0.204	±0.860	±1.728	±0.206	±2.564
	δ n-1 (%)	5.06	28.09	7.49	1.46	6.26	7.20	0.58	7.70
	RSD	403.25	61.01	161.99	624.68	411.14	460.35	411.96	503.40
	Range (%)	0.0 - 51.2	0.0 - 100.0	0.0 - 25.8	0.0 - 14.6	0.0 - 75.0	0.0 - 50.0	0.0 - 2.8	0.0 - 46.8
Illite	n	180	153	83	201	207	70	34	38
	$\bar{X}$ (%)	89.26	50.41	60.29	97.27	63.60	42.68	9.90	26.47
	95% conf. int. (%)	±2.745	±4.457	±4.876	±1.146	±2.465	±3.558	±3.891	±4.642
	δ n-1 (%)	18.61	27.81	22.20	8.22	17.95	14.81	10.99	13.94
	RSD	20.85	55.17	36.82	8.45	28.22	34.71	111.00	52.65
	Range (%)	2.0 - 100.0	0.0 - 100.0	5.2 - 100.0	48.5 - 100.0	8.1 - 100.0	8.3 - 93.5	0.0 - 68.0	0.0 - 74.1
I_{002}/I_{001}	n	165	90	75	192	197	63	5	29
	$\bar{X}$	0.40	0.43	0.39	0.39	0.36	0.34	0.33	0.38
	95% conf. int.	±0.018	±0.031	±0.023	±0.011	±0.014	±0.028	±0.149	±0.028
	δ n-1	0.12	0.15	0.10	0.08	0.10	0.11	0.11	0.07
	RSD	29.11	34.11	25.40	20.08	28.16	32.14	32.57	18.99
	Range	0.23 - 0.88	0.13 - 0.85	0.18 - 0.70	0.17 - 0.75	0.15 - 0.77	0.16 - 0.67	0.19 - 0.45	0.24 - 0.57
Kubler Index	n	174	144	80	199	200	70	25	37
	$\bar{X}$	5.03	8.78	5.02	5.26	5.02	5.45	5.59	4.45
	95% conf. int.	±0.160	±0.496	±0.245	±0.139	±0.151	±0.363	±0.730	±0.346
	δ n-1	1.07	3.00	1.09	0.99	1.08	1.51	1.73	1.03
	RSD	21.20	34.19	21.80	18.80	21.53	27.75	30.99	23.05
	Range	2.6 - 9.0	3.2 - 15.8	3.2 - 8.2	1.8 - 8.2	2.1 - 7.6	2.5 - 10.7	3.0 - 9.9	3.1 - 7.6

where: n = number of samples; δ n-1 = standard deviation; $\bar{X}$ = mean value; RSD = relative standard deviation

Formation (WPb)is identified by a characteristic "kaolinite hump" on the clay mineral logs (Appendix A). The clay mineral characteristics of the Otherside Formation (OF) are incompletely known owing to insufficient drill hole coverage. The base of the formation, however, is marked by another "kaolinite hump" on the logs (Appendix A), identifying the transition from the Locker Lake Formation(LL). In drill holes located within the Carswell structure, the Otherside Formation (OF) is marked by illite predominance and minor chlorite, interspersed by multiple kaolinitic intervals. The Tuma Lake Formation could not be identified as a distict unit in any of the anlyzed drill cores and is here included with the Otherside Formation. Clay-rich sandstones of the Douglas Formation (DG), known only from within the Carswell Structure, contain an assemblage of illite and chlorite, similar to that of the upper Wolverine Point Formation; dolomitic intercalations also contain kaolinite and, consequently, carry an assemblage similar to that of the overlying Carswell Formation (CW) (Hoeve et al., 1985).

Correlation of clay mineral intervals from one drill hole to the next along section B-B" (Figure 1; 4) renders a clay mineral stratigraphy which shows an excellent match with Ramaekers' (1979; 1980; 1981) litho-stratigraphy, except for the Lazenby Lake Formation (Figure 4). According to Ramaekers (op. cit.) the Lazenby Lake Formation comprises a thin (100-150 m) pebbly, marine, sandstone unit, which in the western and central sub-basins overlies the Fair Point and Manitou Falls Formations.

Consistent with this interpretation, the Lazenby Lake Formation is identified in drill holes GE-2 and CSP2-1 (Appendix A) as a 100 m thick, illitic interval between the lower Wolverine Point and upper Manitou

Falls Formations. Drill hole CSP8-1, however, reveals a discrepancy between the litho- and clay mineral stratigraphies. In Ramaekers' (1980) interpretation of this drill hole, the lower Wolverine Point Formation (WPa) is underlain by 100 m of Lazenby Lake Formation, which in turn is underlain by fine-grained sandstones, assigned to the upper Manitou Falls Formation (MFc + MFd). In the clay log of CSP8-1 (Appendix A), the lower Wolverine Point Formation (WPa) is underlain by an illitic clay mineral interval, which near its top contains some kaolinite and further down becomes exclusively illitic. In drill hole CSP10-1 (Section B-B'; Figures 2,3,5), the lower Wolverine Point Formation (WPa) is underlain by a clay mineral interval marked by illite ± kaolinite, which corresponds to Ramaekers' (1980) Lazenby Lake Formation. This unit is in turn underlain by an interval which Ramaekers had assigned to the upper Manitou Falls Formation (MFd), but whose clay mineral assemblage is entirely illitic.

Similar illitic intersections are noted in several other drill holes, not studied by Ramaekers, which are located in the vicinity of the Carswell Structure (Section A-A ; Figures 2,6; Appendix A). In drill hole CAR 114, a 600 m illitic interval underlies the lower Wolverine Point Formation (WPa) and overlies the basement. Drill hole CAR 301, though not reaching basement, displays similar relationships. In AS-1, a 350 m thick illitic interval is found between the lower Wolverine Point Formation (WPa) and the metamorphic basement, whereas AS-2 intersects 400 m of a similar interval before terminating short of the basement. In drill hole BM-1, located 100 km northeast of the Carswell structure, the illitic interval is not present and the lower Wolverine

Point Formation (WPa) directly overlies the upper Manitou Falls Formation (MFc + d).

Interpreting these relationships, the illitic intersections appear to represent a distinct lithological unit of limited areal extent. This notion is supported by the observation of common petrographic characteristics for the rocks within the illitic intervals i.e. fine, uniform or bimodal grain size distribution, high roundness and sphericity, low clay content, absence of detrital micas and heavy mineral grains, abundance of quartz overgrowths, prevalence of point contacts and floating grains. Together these features point to a strong aeolian influence in the depositional area (Turner, 1980). In drill core this unit is marked by a paucity of clay/silt intraclasts, which allows it to be distinguished from the upper Manitou Falls Formation (MFd) with which it might otherwise be confused.

On the basis of clay assemblage and petrographic characteristics, we include the illitic unit with the Lazenby Lake Formation. In this interpretation, Ramaekers' Lazenby Lake Formation is but the uppermost portion of a much thicker lithostratigraphic unit, which is particularly well developed in the general vicinity of the Carswell structure. This new and expanded Lazenby Lake Formation forms a third clastic wedge, stratigraphically equivalent to the Fair Point and Manitou Falls Formations, which enters the basin from the southwest and which distally shows onlapping and interfingering relationships with these two units. This interpretation is consistent with Ramaekers (1979, 1980, 1981) observation of increasing grain size to the southwest, indicating sediment influx from that direction.

Clay/Silt Layers

The intercalated clay/silt layers consist of the same clay minerals, kaolinite, illite and chlorite, as the sandstones and conglomerates; the proportions are vastly different, however. Apatite, goyazite and anatase are noted as accessories. In addition, diaspore was identified in several stratigraphic levels of drill hole CAR 114 (Appendix A).

The clay/silt layers in most of the formations of the Athabasca Group are marked by a monotonous kaolinite-rich assemblage (Table 5), which defies stratigraphic differentiation on the basis of clay mineralogy (Appendix A). Even where the adjoining sandstone is extremely illitic as in the Locker Lake (LL) and Lazenby Lake (LzL) Formations, intercalated clay/silt layers are still primarily kaolinitic in composition. The upper Wolverine Point Formation (WPb) is distinguished by an illitechlorite (sudoite) assemblage similar to that in the sandstones of this formation (Appendix A; Table 2). Although we have insufficient coverage on clay/silt layers within the Douglas Formation, the few samples that we have analyzed suggest similar a illite-chlorite assemblage as within the upper Wolverine Point Formation. No information is available for clay/silt layers in the Otherside and Carswell Formations.

Clay/Silt Intraclasts

The intraclasts are of a similar composition as the intercalated clay/silt layers (Table 6), but are even more kaolinite-rich, except for the upper Wolverine Point Formation which is again distinguished by illite and chlorite (sudoite). Tuffaceous clasts in this formation are primarily composed of fine-grained quartz and K-feldspar or of illite.

Fault-controlled Kaolinization

The regular clay mineral stratigraphy of the sandstones and con-

Table 5. Clay Mineral Data for Athabasca Group Stratigraphic Subdivision

Clay Seam Samples

		LL + OF	WPb	WPa	LzL	MF c+d	MF a+b	FPb	FPa
Kaolinite	n	48	202	59	56	79	26	5	6
	$\bar{x}$ (%)	70.44	8.09	64.03	77.84	86.03	85.69	97.80	84.50
	95% conf. int. (%)	±8.068	±3.074	±7.904	±5.767	±2.796	±4.798	±3.165	±10.580
	δ n-1 (%)	27.50	22.10	30.07	21.34	12.40	11.65	2.28	9.20
	RSD (%)	39.04	273.05	46.97	27.42	14.42	13.59	2.33	10.89
	Range (%)	0.0 - 100.0	0.0 - 95.0	0.0 - 100.0	0.0 - 100.0	61.0 - 100.0	95.0 - 100.0	95.0 - 100.0	69.0 - 93.0
Chlorite	n	48	202	59	56	79	26	5	6
	$\bar{x}$ (%)	0.13	16.98	0.19	1.25	0.00	0.00	0.00	2.33
	95% conf. int. (%)	±0.144	±2.489	±0.192	±1.311	*	*	*	±6.571
	δ n-1 (%)	0.49	17.90	0.73	4.85	*	*	*	5.72
	RSD (%)	391.40	105.45	391.81	388.11	*	*	*	244.95
	Range (%)	0.0 - 3.0	0.0 - 85.0	0.0 - 5.0	0.0 - 26.0	*	*	*	0.0 - 14.0
Illite	n	48	202	59	56	79	26	5	6
	$\bar{x}$ (%)	29.44	74.93	35.78	20.91	13.97	14.31	2.20	13.17
	95% conf. int. (%)	±8.036	±3.288	±7.908	±4.903	±2.796	±4.798	±3.165	±10.345
	δ n-1 (%)	27.39	23.65	30.09	18.15	12.40	11.65	2.28	9.00
	RSD (%)	93.04	31.56	84.09	86.77	88.76	81.41	103.65	68.34
	Range (%)	0.0 - 100.0	5.0 - 100.0	0.0 - 98.0	0.0 - 78.0	0.0 - 55.0	0.0 - 39.0	0.0 - 5.0	7.0 - 31.0
I_{002}/I_{001}	n	37	129	40	38	47	16	0	5
	$\bar{x}$ (%)	0.38	0.31	0.38	0.44	0.41	0.36	*	0.41
	95% conf. int. (%)	±0.040	±0.027	±0.031	±0.036	±0.042	±0.073	*	±0.269
	δ n-1 (%)	0.12	0.16	0.10	0.11	0.14	0.13	*	0.19
	RSD (%)	31.37	50.37	25.24	24.33	34.24	37.41	*	47.57
	Range (%)	0.15 - 0.68	0.08 - 0.92	0.14 - 0.59	0.24 - 0.68	0.18 - 0.75	0.15 - 0.74	*	0.12 - 0.60
Kubler Index	n	42	197	55	52	64	19	0	5
	$\bar{x}$ (%)	5.67	11.78	7.71	4.99	4.56	3.61	*	5.48
	95% conf. int. (%)	±0.664	±0.372	±1.175	±0.583	±0.340	±0.461	*	±2.350
	δ n-1 (%)	2.11	2.64	4.31	2.07	1.35	0.93	*	1.69
	RSD (%)	37.15	22.41	55.90	41.57	29.68	25.81	*	30.90
	Range (%)	2.9 - 15.7	2.8 - 16.7	3.1 - 16.9	1.9 - 13.4	3.0 - 9.7	1.8 - 4.7	*	4.2 - 8.3

Note: * indicates "Not Available"

where: n = number of samples, δ n-1 = standard deviation
$\bar{x}$ = mean value, RSD = relative standard deviation

Table 6. Clay Mineral Data for Athabasca Group Stratigraphic Subdivisions

Clay Intraclast Samples

		LL + OF	WPb	WPa	LzL	MF c+d	MF a+	FPb	FPa
Kaolinite	n	1	16	62	21	128	28	0	1
	$\bar{x}$ (%)	0.00	6.25	84.35	18.71	87.80	89.07	*	78.00
	95% conf. int. (%)	*	±13.761	±2.633	±10.084	±1.444	±4.74	*	*
	δ n-1 (%)	*	25.00	10.29	21.62	8.22	11.96	*	*
	RSD (%)	*	400.00	12.19	115.51	9.36	13.42	*	*
	Range (%)	*	0.0 - 100.00	46.0 - 100.0	0.0 - 79.0	37.0 - 100.0	52.0 - 100.0	*	*
Chlorite	n	1	16	62	21	128	28	0	1
	$\bar{x}$ (%)	0.00	23.69	0.00	0.00	0.13	0.00	*	0.00
	95% conf. int. (%)	*	±16.124	*	*	±0.187	*	*	*
	δ n-1 (%)	*	29.29	*	*	1.07	*	*	*
	RSD (%)	*	123.67	*	*	851.86	*	*	*
	Range (%)	*	0.0 - 99.0	*	*	0.0 - 11.0	*	*	*
Illite	n	1	16	62	21	128	28	0	1
	$\bar{x}$ (%)	100.00	70.06	15.65	81.29	12.16	10.93	*	22.00
	95% conf. int. (%)	*	±18.805	±2.633	±10.084	±1.432	±4.74	*	*
	δ n-1 (%)	*	34.17	10.285	21.62	8.15	11.96	*	*
	RSD (%)	*	48.76	65.74	26.59	67.07	109.41	*	*
	Range (%)	*	0.0 - 100.0	0.0 - 54.0	21.0 - 100.0	0.0 - 63.0	0.0 - 48.0	*	*
I_{002}/I_{001}	n	1	10	49	21	100	13	0	1
	$\bar{x}$ (%)	0.47	0.26	0.49	0.49	0.53	0.47	*	0.48
	95% conf. int. (%)	*	±0.068	±0.030	±0.068	±0.026	±0.063	*	*
	δ n-1 (%)	*	0.09	0.10	0.15	0.13	0.10	*	*
	RSD (%)	*	35.23	21.42	29.87	24.10	21.18	*	*
	Range (%)	*	0.12 - 0.36	0.31 - 0.81	0.32 - 0.84	0.25 - 0.89	0.37 - 0.68	*	*
Kubler Index	n	1	13	57	21	115	20	0	1
	$\bar{x}$ (%)	4.4	11.15	3.78	5.37	3.39	3.32	*	5.7
	95% conf. int. (%)	*	±1.325	±0.322	±0.502	±0.267	±0.319	*	*
	δ n-1 (%)	*	2.11	1.20	1.08	1.44	0.66	*	*
	RSD (%)	*	18.88	31.79	20.04	42.38	19.99	*	*
	Range (%)	*	7.3 - 13.8	2.2 - 6.9	4.1 - 8.4	0.8 - 10.1	1.9 - 4.4	*	*

Note: * indicates "Not Available"

where: n = number of samples, δ n-1 = standard deviation,
$\bar{x}$ = mean value, RSD = relative standard deviation

glomerates is locally transected by fault-controlled zones of kaolinitization. Although a localized feature, this kaolinitization tends to be pervasive, leading to total obliteration of the normal diagenetic clay mineral assemblages. It is accompanied by a decrease in total clay content to approximately 1%, from 5-20% in unaffected sandstones, and by a loss of coherence, which ultimately leads to the formation of an almost clay-free, loose sand. Apparently, the kaolinitization represents a process of intensive leaching which destroyed most of the interstitial clay cement and transformed the remainder into kaolinite.

An example is provided by drill core NR-1, located near the extrapolated extension of the Grease River Fault zone (Figures 1,2,3). This drill hole intersects the Locker Lake and Otherside Formations (Ramaekers, 1979; 1980), whose normal diagenetic clay mineral assemblage is predominantly illite; yet the clay mineral log for NR-1 shows extreme kaolinite (Appendix A). A similar situation is observed in drill core CSP 6-1 (Figures 2,3; clay log not shown) and at various other localities within the basin. On a much smaller scale, fracture zones intersected in drill core are frequently flanked by kaolinite haloes in the order of metres or tens of metres wide (Appendix A; AS-1).

Relying on company information, we had previously interpreted the kaolinite-rich shoulder on the clay mineral log for drill hole AS-1 (Appendix A) as an interval of Fair Point Formation (Hoeve et al., 1985).

Since then, however, we have examined the drill core and realized that the kaolinite-rich interval represents a zone of intensive fracturing and late kaolinitization. Therefore, no Fair Point Formation is present

in drill hole AS-1 and the eastern limit of this formation is located to the west of what we had indicated in our earlier publication.

Clay Mineral Characteristics

The chlorites of the Athabasca Group sediments are of the di,trioctahedral variety (Sudoite; nomenclature after Bailey, 1980 a), characterized by d(060) ~ 1.50 Å and by a very high intensity of the (003) reflection. Regular trioctahedral chlorites have not been found anywhere in the sedimentary sequence. The characteristics of illite are variable and are discussed below.

Illite, "crystallinity" and I(002)/I(001) ratio

Weaver (1956; 1960) and Kübler (1964) have shown that the "crystallinity" of illite, measured as the width of the (001) peak at half height, progressively increases with depth of burial, and Kübler (1968) utilized this feature to bracket the anchizone between high-grade diagenesis and low-grade metamorphism. Subsequently, Esquevin (1969) has refined Kübler's technique of evaluating diagenetic grade on the basis of illite "crystallinity." Plotting Kübler's index against the intensity ratio of the (002) and (001) XRD reflections, which he considered proportional to the compositional ratio Al/Mg + Fe of the octahedral layer, Esquevin demonstrated that the recrystallization of illite also depends on compositional factors. He proposed that aluminary illites with I(002)/I(001) > 0.26, contrary to magnesian illites, would be useful indicators of the diagenetic conditions a sediment has experienced. Dunoyer de Segonzac (1969) made extensive use of Esquevin diagrams in studies of the burial history of several sedimentary basins in Europe and north Africa. On the basis of illites plotting within distinct fields on the diagram, he differentiates between five environments

of recrystallization ranging from incipient diagenesis to low-grade metamorphism (Figure 7a).

I. Incipient diagenesis. Very weak recrystallization, the illites are of a highly variable crystallinity representing a mixture of authigenic and detrital material.

II. Low-grade diagenesis. Weak to moderate recrystallization; illite crystallinity remains variable.

III. High-grade diagenesis. Strong recrystallization; illite crystallinity values rather uniform within the range Ki = 4 - 7.

IV. Anchizone. Transition to metamorphism. Illite crystallinity is independent of lithology; values within the range Ki = 2.5 - 4.0.

V. Low-grade metamorphism. Illites are progressively transformed into white micas. Crystallinity values range within narrow limits around Ki = 2.5, regardless of I(002)/I(001) ratios and composition.

Kübler's and Esquevin's use of illite "crystallinity" and (002)/(001) peak intensity ratio as indices of diagenetic grade has been criticized by other authors who argued that the values of these parameters is also influenced by additional compositional factors (Weaver, 1965; Weaver and Pollard, 1973). Srodon (1984) and Weaver and Broekstra, (1984) argued that both the (001) peak width and the (002)/(001) peak intensity ratio largely depend on the proportion and composition of intercalated smectite layers. Nonetheless, a correlation between illite "crystallinity", (002)/(001) peak intensity ratio and depth of burial has been documented in many studies (Dunoyer de Segonzac, 1969; 1970; many references in Kish, 1983), so that Esquevin plots provide a convenient empirical method to evaluate diagenetic progress.

For the Athabasca Group rocks, Kübler indices for illites of sandstones as well as intercalated clay/silt layers and intraclasts, are rather uniform, averaging 4-5 (Appendix A; Tables 1, 2, 3). The Upper Wolverine Point and Douglas Formations are the only stratigraphic units that have much higher values of approximately K_i = 8 (sandstone) and K_i = 12 (clay/silt), respectively.

Esquevin diagrams for sandstones and conglomerates of the lower Manitou Falls Formation (MFa + b; Figure 8b) show a cluster of data points characteristic of Dunoyer de Segonzac's field (III) of high grade diagenesis (Figure 7a). The diagrams for the other formations (not shown) are similar, with the exception of the upper Wolverine Point and Douglas Formations (Figure 7c), which display a broad scatter of data points characteristic of Dunoyer de Segonzac's field (I), suggesting weak diagenetic recrystallization. Similar, but more pronounced differences between Manitou Falls and upper Wolverine Point Formations are apparent in the diagrams for the silt/shale sets of samples (Figure 7d, 7e).

Intraformational differences between the sandstone and clay/silt samples are minor, except for the upper Wolverine Point Formation, where the clay/silt samples have a much higher Kübler index (Ki = 12 as opposed to Ki = 8 - 9) and a much lower value of the illite peak intensity ratio I(002)/I(001), pointing to a poorer "crystallinity" and a different composition (Appendix A; Tables 1,2,3). For the Douglas Formation insufficient data are available to make a similar comparison.

Smectite interlayering in illite

With increasing depth of burial smectites are progressively converted into illite via random and regularly ordered I/S intermediate phases. This process was first documented for the Texas Gulf Coast (Burst, 1959) and has since been studied in many other sedimentary basins (cf. Kish, 1983). It appears that the smectite/illite evolution depends not only on temperature, but also on residence time and on potassium availability. For this reason I/S interlayering cannot be used as a geothermometer (Srodon and Eberl, 1984). Concomitantly with increasing proportions of illite component the degree of ordering progresses from random to ordered IS and ISII structures. The transition from random to IS ordering takes place at aproximately 45% smectite whereas ISII ordered structures appear at approximately 15% smectite.

At the Texas Gulf Coast, the smectite evolution is arrested at approximately T = 100°C with an I/S of 20% smectite and a IS ordered structure, (Perry and Hower, 1970; Hower et al., 1976; Boles and Franks, 1979) whereas in other basins the evolution proceeds towards illite through ISII ordered interlayering (Hoffman and Hower, 1979). Once the ISII ordered structure is reached, however, the conversion slows down and the remaining < 10% smectite layers prove to be very persistent. Weaver and Broekstra (1984) report 10% smectite layers at the transition from the diagenesic zone to the anchizone and the disappearance of the last expandable layers at the beginning of the epizone.

For the Athabasca Group rocks, selected illite-rich samples were tested for the presence of expandable phases by means of treatment with ethylene glycol. In samples of the upper Wolverine Point Formation, the illites contain an expandable component, whose proportion is estimated

at 10-15%, using methods and criteria developed by Srodon (1984). Lithology seems to be a factor, however, because the proportion of smectite layers appears to be somewhat higher in samples of clay/silt layers and intraclasts than in those of adjoining sandstones of this formation. By contrast, the illites of the other formations are pure illites. In those formations smectites and I/S mixed-layer phases are conspicuously absent.

These differences in illite characteristics are unlikely to reflect different diagenetic conditions because the Wolverine Point Formation must have been subjected to essentially the same temperatures as the other formations, nor can they be due to differences in potassium availability because the Wolverine Point Formation contains an immature detrital assemblage relatively rich in potassium carrier minerals. More likely the differences reflect primary lithological differences and different starting materials.

Illite polytypes

With increasing depth of burial illite polytypes change from 1 M_d to 1 M to 2 M. Increasing 2 M/1 M_d ratios with increasing burial metamorphic grade have been documented in many sedimentary basins (c.f. review by Kish, 1983). According to Srodon and Eberl (1984) the end product of the diagenetic evolution of smectite is an illite of 1 M polytype. This same 1 M polytype apparently also characterizes illites that formed by direct replacement of kaolinite.

For Athabasca Group rocks, illites of the Wolverine Point Formation were identified as to consist of 1 M_d/1 M mixtures, based on criteria developed by Srodon and Eberl (1984) and XRD data listed by Bailey (1984 b) and Carroll (1970). In clay/silt layers and intraclast the polytype

is primarily 1 Md, whereas in adjoining sandstones of this formation are characterized by predominantly 1 M. The illites of the other formations consist of mixtures of 1 M and 2 M, in variable proportions, judging from the intensities of diagnostic XRD reflections.

PROGRADE AND RETROGRADE DIAGENESIS

Clay mineral characteristics of the analyzed samples indicate that the Athabasca Group rocks have been subjected to high-grade diagenesis. This is attested to, first of all, by the (1) presence of dickite fracture fillings; (2) illites being of mixtures of 1 M/2 M polytype and lacking expandable layers in most of the sandstones; (3) illites of the upper Wolverine Point Formation containing $\leq$ 15-10% expandable layers; (Maxwell and Hower, 1967; Dunoyer de Segonzac, 1969, 1970; Hoffman and Hower, 1981; Kish, 1983; Srodon, 1984; Srodon and Eberl, 1984; Weaver and Broekstra, 1984). Furthermore, the illite crystallinity values of 4-5 are characteristic of those in the lower portion of a deep sedimentary basin, as compared to values of 15 or more for sediments that have not been deeply buried (Kübler, 1964, 1966; Dunoyer de Segonzac, 1969, 1970). The influence of depth of burial is illustrated in deep intersections such as the Rumple Lake drill core in which the Kübler index shows an increase from 4-5 at the bottom to 6-7 at the top, whereas the influence of lithology is apparent for the upper Wolverine Point and Douglas Formations (Appendix A). It is clear that the Athabasca Group forms the lower portion of a deeply buried sedimentary sequence whose thickness must have been much in excess of the presently preserved 2000 m. Pagel's (1975a; 1975b) estimate of 4-5 km on the basis of fluid inclusion studies seems reasonable.

Esquevin plots for the sandstones and conglomerates of the Manitou Falls Formation are characteristic of Dunoyer de Segonzac's (1969; 1970) zone of high-grade diagenesis suggesting temperatures in the order of 150-200°C. Again, the clay mineral data are in agreement with Pagel's

(1975a; 1975b) estimate of 200°C on the basis of fluid inclusion studies.

The original clay mineral composition of the Athabasca Group rocks and the effects of diagenesis can be evaluated by comparing the clay mineral assemblages of sandstones with those of intercalated clay/silt layers and intraclasts.

On the assumption that the provenance areas were underlain by a kaolinitic weathering profile similar to that preserved in the floor of the basin, kaolinite may be expected to have been the main detrital clay mineral that was carried into the basin. This assumption is supported by the fact that intercalated clay/silt layers and intraclasts, in units other than the upper Wolverine Point (WPb) and Douglas (DG) Formations, exhibit a uniform, extremely kaolinite-rich composition, irrespective of lithology or stratigraphy. The same applies for the >2μm fraction of sandstones and conglomerates which consists largely of kaolinite. By contrast, a coherent clay mineral stratigraphy which closely matches the lithostratigraphy is observed in the ⩽2μm fraction of sandstones and conglomerates. These relationships indicate that in the sandstones the illites of the present illite-kaolinite assemblages have been produced during diagenesis by conversion of kaolinite into illite, a conclusion which is supported by petrographic evidence indicating replacement of kaolinite by illite.

The fact that the clay mineral stratigraphy closely matches the lithostratigraphy means that the present proportions of illite and kaolinite in the ⩽2μm fraction must have been governed by lithologically controlled potassium availability. Considering that petrographic evidence points to intensive post-depositional alteration, detrital mica

and feldspar may have provided a source of potassium during diagenesis. Nonetheless, as a significant proportion of kaolinite is still preserved within the sandstones and conglomerates of the Manitou Falls and Fair Point Formations, these detrital minerals have probably constituted an insufficient reservoir of potassium for the conversion to proceed to completion. Estimating the average clay content of the Manitou Falls Formation at approximately 10%, a few percent of detrital feldspar could have released sufficient potassium to convert all detrital kaolinite into illite. Thus, the partial preservation of kaolinite sets an upper limit in the order of 1 percent to the abundance of detrital feldspar and mica in the original sediments, an estimate which is consistent with petrographic evidence showing a general paucity of clay pseudomorphs after these minerals.

In view of the foregoing, we disagree with Ramaekers' (1979, 1980, 1981) assertion that virtually all of the interstitial clay was derived from intrastratal weathering of detrital grains. It is our interpretation that, in units other than the Wolverine Point and Douglas Formations, virtually all clay derives from detrital kaolinite or a halloysitic precursor, which to a variable degree was directly converted into illite, depending on lithologically controlled potassium availability. Post-depositional alteration of mica and feldspar, though having certainly taken place, only made an insignificant contribution towards clay formation owing to a general lack of potassium-bearing detrital minerals. Direct replacement of kaolinite by illite is a process that is well-known from the literature (see numerous references in Kish, 1983) and accounts for the disappearance of kaolinite with increasing depth of burial, as documented by Dunoyer de Segonzac, 1970). Contrary to the

illitization of smectites, this process is not strongly dependant on temperature and is generally completed well before the disappearance of the last expandable layers in illite derived from smectite. Weaver and Broekstra, (1984) and Srodon and Eberl (1984) found that small amounts of smectite layers in illite persist until the anchizone/epizone transition, whereas kaolinite generally disappears under diagenetic conditions of much lower grade (Dunoyer de Segonzac, 1970).

Similarly we contend that intrastratal oxidation of detrital ferromagnesian minerals has been of little importance to the development of the hematite pigment. We propose that most of the dispersed hematite has been derived from colloidal limonite, or other hematite precursors, which were carried into the basin along with detrital quartz, kaolinite and accessory detrital Fe-Ti-oxides. Sediments derived from modern lateritic terranes carry abundant yellow and brown iron hydroxides, which upon diagenesis convert into hematite and impart a red colour to the rocks (Van Houten, 1968; Walker, 1974; Riezebos, 1979).

The upper Wolverine Point (WPb) and Douglas (DG) Formations differ from the other formations by their clay mineral assemblage consisting of illite and chlorite (sudoite), which points to a different sediment source, in agreement with Ramaekers' (1980, 1981) interpretation of a marine, inner to open shelf, depositional environment. The fact that illites of the upper Wolverine Point Formation still contain ~ 10% expandable layers suggests that they were derived from smectite. By analogy with comparable more recent and modern phosphatic and glauconitic sediments, the detrital clay mineral assemblage of the upper Wolverine Point Formation may have been dominated by smectites, mixed-layer minerals, poorly ordered illites and glauconite (Weaver and Wampler,

1972; Riggs, 1979a,b; Balson, 1980; Lucas et al., 1980). The potassium and magnesium required to convert such an association into the present illite-chlorite assemblage may have been derived from detrital feldspar, ferromagnesian minerals and tuffaceous components, because petrographic observations point to an immature original sediment rich in such constituents. The proportions of illite and chlorite in intercalated clay/silt layers and intraclasts are similar, illustrating that their compositions have not been differentially modified during diagenesis. The sandstones, however, are relatively richer in chlorite, reflecting preferential formation of this mineral.

Esquevin diagrams for the upper Wolverine Point Formation are characteristic of Dunoyer de Segonzac's field I of early diagenesis (Figure 7a,c). Taking into account that all preserved stratigraphic units of the Athabasca Group must have been subjected to essentially the same condition of deep burial and high-grade diagenesis, the apparent lower degree of diagenesis in the upper Wolverine Point Formation may be attributed to lithological factors. The presence of up to 15% of expandable layers in the illites of this formation, reflecting a different diagenetic evolution as compared to the kaolinite-rich sandstones of the other formations, may account for the higher values of the K bler index and the broader range of (002)/(001) peak intensity ratios (Appendix A; Figure 7c, e). A low permeability during diagenesis due to abundant clay, silt, phosphate and tuff, may have been an important additional factor, as permeable sandstones are known to be more reactive than their cemented counterparts and shales (Dunoyer de Segonzac, 1969; 1970; Kish, 1983). Early reduction of porosity is also suggested by a Rb/Sr-dating study by Armstrong and Ramaekers (1985), which indicates that Rb/Sr and

radiogenic Sr in shale/silt-rich Wolverine Point Formation had not been homogenized during diagenesis. Permeability effects may also be responsible for apparent differences in the degree of diagenesis reached by sandstones of the Wolverine Point Formation on the one hand and clay/silt layers and intraclasts on the other, as expressed by different Esquevin plots (Figure 7 c, e), different proportions of expandable layers in illite and in polytype.

Whilst the features discussed so far concern mineralogical changes during prograde burial and diagenesis, retrograde diagenetic changes are exemplified by late fault-controlled kaolinitization. Similar late kaolinitization has been reported for other sedimentary basins where it has been interpreted as related to uplift and unroofing, during which acidic surface waters percolating downwards along reactivated open fault zones, regained access to the deep aquifer (Dunoyer de Segonzac, 1969; 1970; Kish, 1983).

It appears from this discussion that two trends, reflecting differences in original detrital mineralogy and sedimentary environment, can be recognized in the prograde diagenetic history of the clay minerals in the Athabasca Group rocks. A continental trend, expressed in Manitou Falls, Lazenby Lake, Fair Point, lower Wolverine Point, Locker Lake and Otherside Formations, started with detrital kaolinite which during burial was recrystallized and variably converted into illite, depending on lithologically-controlled potassium availability. A marine trend in the upper Wolverine Point and Douglas Formations, probably started with a detrital association of smectite, poorly ordered illite and smectite-illite mixed-layer minerals, which during burial was transformed into a diagenetic illite-chlorite assemblage.

This interpretation, which calls for a closed-system behaviour of the major lithological units during progressive burial, is supported by the observation that the clay mineral stratigraphy of the sandstones and conglomerates closely matches the litho-stratigraphy. The preservation of such relationships implies absence of significant cross-formational flow of pore waters between units with contrasting clay mineral assemblages e.g. Wolverine Point versus Manitou Falls, Lazenby Lake and Fair Point Formations.

DEPOSITIONAL ENVIRONMENT AND STRATIGRAPHIC DEVELOPMENT

Our investigation of clay mineral distribution and diagenetic patterns casts doubt on Ramaeker's (1979; 1980; 1981) interpretation that the Athabasca Group rocks, except those of the Manitou Falls Formation, were deposited in a marine environment. The recognition of kaolinite as the almost exclusive detrital clay mineral in the Manitou Falls, Lazenby Lake, Fair Point, lower Wolverine Point and Otherside Formation points to a terrestrial depositional environment. This is supported by the identification of diaspore at several stratigraphic levels (drill hole CAR 114; Figure 4). Diaspore was found in the sub-Athabasca palaeo-weathering profile and at several levels within the overlying sandstones. Most of these occurrences are within the Lazenby Lake Formation and one is within the Locker Lake Formation. The identification of the mineral within the palaeo-weathering profile confirms earlier interpretations as to the lateritic nature of this profile (Hoeve and Sibbald, 1978; MacDonald, 1980). The presence of diaspore at different levels within the sandstones may suggest intervals of nondeposition and intensive weathering or, alternatively, the diaspore may represent washed-in material derived from a nearby weathering profile. In either case, the mineral is indicative of a fluviatile terrestrial rather than a marine depositional environment for the Lazenby Lake and Locker Lake Formations. This is consistent with the observation of sporadic granules and pebbles of phosphate in the Locker Lake Formation, which may indicate erosion of the underlying Wolverine Point Formation.

The Lazenby Lake Formation was identified in this study as a third clastic wedge, stratigraphically equivalent to the Manitou Falls and Fair Point Formations. Based upon the areal extent of the Lazenby Lake

Formation, and in accordance with a seismic study by Hobson and McAulay (1969), we have modified Ramaekers' (1980, 1981) configuration of the Mirror sub-basin (Figure 3). The Patterson High, instead of forming a broad high separating the Jackfish and Mirror sub-basins, is reduced to a ridge which divides the latter into a western and an eastern arm. The Lazenby Lake Formation was initially restricted to the western arm and it was only after the Patterson High had been overcome that the formation began to spread out eastwards to interact with the approaching clastics of the upper Manitou Falls Formation (MFc + d). Towards the west and north it interfingers with and onlaps onto the Fair Point and Manitou Falls Formations.

In view of these relationships we interpret the Athabasca Group to consist of an essentially fluviatile sequence interspersed by two marine incursions (Table 2). During early stages of basin development, the Jackfish, Mirror and Cree sub-basins were filled in with terrestrial detritus entering the basin from all directions except north and northwest, because the Manitou Falls, Lazenby Lake and Fair Point form fluviatile wedges that project into the basin from the northeast to southeast, the south and the southwest respectively. Towards the end of this period the upper Manitou Falls Formation (MFc + d) becomes the predominant system. Drainage must have been to the northwest.

The first marine incursion is represented by the Wolverine Point Formation, whose lower member (WPa) may represent the transgressive phase of the incursion since it is in this unit that chlorite begins to regularly appear in the clay mineral assemblage, possibly reflecting a modest influx of marine clays. Clay/silt layers and intraclasts, how-

ever, are still extremely kaolinitic, pointing to a continued strong terrestrial influence.

The overlying Locker Lake and Otherside Formations represent a return to fluviatile, terrestrial conditions, which is followed by the second marine incursion embodied by the illite/chlorite-rich shales and sandstones of the Douglas Formation and the stromatolitic dolomites of the Carswell Formation. Marine conditions may subsequently have prevailed, as indicated in Middle- and Late-Proterozoic stratigraphic and tectonic reconstructions of Western Canada (Young, 1984). In the Athabasca basin, however, the Carswell Formation marks the end of the preserved stratigraphic record.

The depositional environment we envisage for the Athabasca Group is that of an intra-continental basin, which was filled in with fluviatile terrestrial sandstones and which had only intermittent contact with the open sea. This contrasts with the unstable continental-margin setting proposed by Ramaekers (1980; 1981), which is characterized by a prograding fluviatile wedge in the east (MF) and a thick sequence of nearshore to shallow-shelf, marine sediments in the west, comprising several transgressive-regressive cycles.

The presence of tuff fragments in intercalated silts and shales at basal, intermediate and upper stratigraphic levels attests to an influx of volcaniclastic material throughout the sedimentary sequence. The absence of such fragments from the sandstones may reflect poor preservation in high energy fluviatile environments, or destruction during diagenesis of permeable sandstones.

URANIUM IN NORTHERN SASKATCHEWAN

All significant uranium mineralization in Northern Saskatchewan is located to the west of the Needle Falls Shear zone (Figure 1), i.e., is confined to the cratonic block west of the lower-Proterozoic continental margin (Lewry et al., 1981). Syngenetic occurrences are represented by a variety of pegmatitic showings which are associated with Hudsonian granitoid intrusions and with metamorphic segregations at certain stratigraphic levels of the lower-Proterozoic supracrustals. Epigenetic deposits of commercial interest comprise (1) vein-type deposits at Beaverlodge, (2) unconformity-type deposits in the Athabasca basin and (3) intra-Athabasca occurrences at higher stratigraphic levels in the Athabasca Group sediments.

These three types of epigenetic deposits formed during three distinct episodes of uranium mineralization and remobilization, whereby later episodes over-printed earlier ones (Table 7; Hoeve et al., 1981a).

Episode 1: $\sim$ 1750-1700 Ma.

The first episode gave rise to Hudsonian vein-type deposits of simple mineralogy. Mineralization is restricted to so-called "linear belts", zones of Hudsonian reworking, that transect the Western Craton. Although most of the deposits are located within the Beaverlodge belt, similar mineralization is found as far east as Black Lake (Figure 1). The ore occurs as veins and fracture fillings of a simple paragenesis comprising pitchblende, brannerite and coffinite, accompanied by minor pyrite, chalcopyrite and galena. Host rock alteration is marked by hematitization and albitization, resulting from fracture-controlled soda metasomatism at a regional scale. The deposits bear the characteristics

Table 7. Summary of published dates for uranium mineralization in the Athabasca basin; compiled from Baadsgaard et al., 1984; Bell, 1981; 1985; Clark and Fogwill, 1985; Clark and Homeniuk, 1982; Cumming et al., 1984; Fryer and Taylor, 1984; Hall et al. in press; Hoeve et al., 1981; Jones, 1980; Koeppel, 1968; Tremblay, 1982; Trocki et al., 1984; Worden et al., 1985.

	Beaverlodge	Rabbit Lake	Collins Bay	Cluff Lake	McLean Lake	Key Lake	Dawn Lake	Midwest Lake	Maurice Bay	Johnson Island	Stewart Island	Fond-du-Lac	Raven/Horseshoe
Beaverlodge vein	1750 - 25 Ma												
Unconformity Stage 1		1345 ± 26 1281 ± 11 1278 ± 21	1384 - 1238 1281 ± 80 † 1296 ± 34 *	1330 ± 30* 1350 1293 ± 36 *	1318 ± 4* 1170 ± 3*	1351 ± 4 1270 1242 ± 1 1233 ± 7	n.d	1331 ± 8 1328 ± 17 1337 ± 11* 1220 *					
Unconformity Stage 2	1100	1085 1078 ± 26	n.d.	1050 ± 20* 1050	n.d.		1100	1110 ± 28 1094 ± 14			1100 - 1200?		
				995 - 945 850 ± 5 800*		918 907 ± 4		700	760				
Remobilization	270	250 215 ± 40 211 ± 24	334 ± 21	380 ± 5 250 ± 20 234*	n.d	351 ± 23 300 ± 6 270 251 ± 21	n.d.	265 ± 50	290	220	318 220	220 215	220

+ - Sm/Nd method; * - K/Ar or Ar^{40}/Ar^{39} method;
. - Pb/Pb method; no symbol - U/Pb method.

of the Na-U-Ti class of metamorphic-hydrothermal uranium deposits (Hoeve, 1982).

The mineralization, dated 1740 ± 25 Ma, is coeval with the second peak of Hudsonian magmatic activity in northern Saskatchewan (Bell, 1981), as represented by the intrusion of late-tectonic granites in the root zones of linear belts exposed farther to the south i.e. the Clearwater and Junction granites in the Clearwater and Virgin River Domains, respectively (Figure 1).

Episode 2: ∿ 1350-1000 Ma.

The second, and most important, episode of uranium metallogenesis generated the unconformity-type deposits in the Athabasca basin. Primary mineralization took place well after deposition of the Athabasca Group rocks, dated at 1450 ± 30 Ma. A first pulse of mineralization at 1330-1250 Ma was followed by a second pulse at 1100-1050 Ma and possibly by an ill-defined and yet-to-be-confirmed pulse at 900-700 Ma (Table 7). The entire episode is coeval with a period of diabase intrusive activity spanning the same 1350-1000 Ma interval.

At Beaverlodge, this episode overprinted the pre-existing Hudsonian mineralization and also led to the formation of new deposits with a complex mineralogy dated at ∿ 1100 Ma (Koeppel, 1968). The Nicholson and Fish Hook Bay deposits are examples of the latter (Hoeve, 1978; 1982c; Sibbald, 1982; Figure 1).

Episode 3: 300-250 Ma.

This episode is recognized as a remobilization event which in unconformity-type deposits gave rise to a paragenesis of sooty pitchblende, coffinite and minor sulphide. The remobilization was particularly strong at the Key Lake deposit, where the Athabasca portion of the

ore body has been intensively reworked (Trocki et al., 1984). It has, however, also been identified in other unconformity-type deposits throughout the basin. At Beaverlodge it is represented by a 270 ± 20 Ma (Koeppel, 1968) paragenesis of pitchblende intimately intergrown with hematite.

Intra-Athabasca mineral occurrences at Maurice Bay, Stewart Island, Johnson Island, Fond-du-Lac, Isle-Brochet and at several locations within the interior of the basin, as well as the basement-hosted Raven and Horseshoe deposits (Figure 1), seem to have been largely shaped during this episode. These occurrences are marked by an intimate association of pitchblende, earthy hematite, goethite and locally intensive silicification, and are not obviously related to the unconformity. It is uncertain whether they present entirely new deposits or have been derived from unconformity-type (?) precursor deposits.

Because this episode is not associated with any magmatic or metamorphic activity, it most probably reflects a period of epeirogenic uplift of the Athabasca basin.

GENERAL CHARACTERISTICS OF UNCONFORMITY-TYPE DEPOSITS

The major unconformity-type uranium deposits in the Athabasca basin, as described by various authors, share a common geological setting (Dahlkamp, 1978; Hoeve and Sibbald, 1976, 1978; Hoeve et al., 1980; Clark and Fogwill, 1985; Jones, 1980; Jones and Heine, 1985; Wallis et al., 1984; Wray and Ayres, 1985). The deposits straddle the unconformity between the Aphebian/Archean metamorphosed basement and the overlying Athabasca Group sandstones and conglomerates. They generally consist of several individual ore bodies which are arranged as beads in a string, separated by intervening barren areas, and are spread out for several kilometres along the strike of steeply dipping, NNE-SSW trending fault zones. The fault zones are in the order of a few to tens of kilometres long and no more than approximately one hundred metres wide. The fault zones off-set the unconformity by up to 50 m and tend to follow the strike of the basement rocks, coinciding, at the level of the unconformity, with the trace of graphitic metapelites in the basement. In the Key Lake, McClean Lake, Collins Bay B and Midwest deposits, mineralization is associated with basal Aphebian pelites, whereas the Rabbit Lake and Dawn Lake deposits are associated with graphite-bearing, dolomitic marbles higher in the lower-Proterozoic stratigraphic sequence.

Host rock alteration forms a distinctive aspect of the deposits and is marked by extensive development of clay minerals, tourmaline (dravite), complicated hematitization/bleaching phenomena, formation of hydrocarbons and destruction of graphite, which affect basement rocks as well as the overlying sandstones. In the case of the Midwest deposit, diabase dikes in basement and sandstone are also altered.

Individual ore bodies tend to have extreme longitudinal shapes, length/width ratios being in the order of 3 to 20. Ore is found both below and above the unconformity, although the proportions between basement and sandstone ore are highly variable from one deposit to the next. At the Collins Bay B, Midwest and McClean deposits, virtually all ore is found at the unconformity, whereas at Key Lake most is in the basement. At Rabbit Lake and Eagle Point, erosion of sandstone has left a record of mineralization in basement only. The mineralization is structurally controlled, occurring as veins, breccia fillings and massive replacements. The sandstone above the ore bodies is generally fractured, or even brecciated, and transected by open veins containing euhedral quartz, tourmaline and base metal sulphides.

Typical ore bodies consist of a near-massive, high-grade core at the unconformity and a lower grade envelope, which extends along faults for a few hundred metres below and above the unconformity. The core and envelope display distinct mineralogical differences, with arsenides featuring prominently in the former and sulphides in the latter.

Mineralization post-dates the deposition of the Athabasca Group by 100-150 Ma. Primary mineralization at 1350-1250 Ma is followed by a second surge of mineral precipitation at 1100-1050 Ma (Table 7) and by later remobilization. Clay mineral host rock alteration is coeval with primary pitchblende mineralization and with illite diagenesis of the sandstones (tables 1, 7). Although the overall episode of mineralization and associated host rock alteration is coeval with the 1350-1000 Ma period of diabase magmatism, there is no direct connection between specific deposits and individual diabase dikes.

ORE PARAGENETIC RELATIONSHIPS

Despite mineralogical and paragenetic differences between individual deposits, the relationships between ore and host rock as well as the overall paragenetic development conform to a remarkably uniform pattern in all deposits studied throughout the basin. The following generalized account is based upon studies of the Midwest, Rabbit Lake and Collins Bay B deposits and on an interpretation of literature descriptions of the McClean and Dawn Lake deposits (Clark and Fogwill, 1985; Wallis et al., 1984).

The ore bodies of the Midwest, Collins Bay B and McClean Lake deposits consist of a near-massive, high-grade core at the unconformity, which is surrounded by lower-grade, more dispersed, fracture-controlled mineralization in sandstone and basement rocks. The high-grade ore forms flat-lying lensoid concentrations which are enclosed in an envelope, up to a few metres thick, consisting of a clayey rock that lacks diagnostic textures of either basement or sandstone. This ore consists of near-massive pitchblende/coffinite and Ni-Co-arsenides, often forming dendritic and nodular aggregates in a laminated or structureless clay matrix composed of illite and di,trioctahedral chlorite (sudoite). In the Collins Bay B deposit the clay matrix also contains dark olive green chamosite (nomenclature after Bailey, 1981). Although similar high-grade near-massive ore concentrations embedded in a clay matrix are also found in basement rocks up to several tens of metres below the unconformity, most of the mineralization in the basement occurs as fine-grained impregnations and fracture fillings in clayey, heavily altered rocks. At Key Lake where most of the ore is below the unconformity, the basement ore seems to be similar to the unconformity ore of the other depos-

Table 8. Idealized paragenetic scheme for the first and second stages of mineralization in unconformity-type uranium deposits of the Athabasca basin

Mineral	Stage 1	Stage 2
Tourmaline (dravite)		—
Quartz		—
Calcite/dolomite		—
Siderite		—
Pitchblende I (lustrous)	—	
Pitchblende II (dull)		—
Co-triarsenides	—	
Ni-diarsenides	—	
Ni-monarsenides	—	
Ni-Co-Fe-sulpharsenides	—	
Ni-Co-Fe sulphides	—	
Fe-Pb-Zn-sulphides		- - - —
Hematite	—	
Goethite		—
Fe^{2+}-chlorite (Berthierine = 7 Å chlorite)		———

its. A related situation is found at Rabbit Lake, where primary mineralization in basement rocks is accompanied by trace quantities of Ni-Co-arsenides.

Pitchblende is generally present in at least three distinct generations. At the Midwest deposit, the first generation consists of a radial-textured, botryoidal variety with high reflectance which displays a tendency towards development of euhedral margins or even may form single euhedral grains; the second comprises a more dull variety which occurs as rims around and replacements of the first variety and also forms fine-grained intergrowths with arsenides and sulphides, as well as massive and colloform aggregates; the third and latest generation is represented by a fine-grained sooty variety, which is associated with sooty coffinite and forms coatings and associated impregnations along fractures and shears in both basement and sandstone.

Paragenetic relationships within the high-grade ore and between ore and clay matrix are extremely complicated (table 8), not the least by the fact that the constituent minerals are present in different types and generations and occur intermittently at several stages in the crystallization sequence, giving rise to apparent reversals and contradictory replacement relationships. A generalized paragenetic succession for the Midwest deposit is as follows: lustrous pitchblende plus illite and di,trioctahedral chlorite (sudoite) are followed by coffinite associated with initially Co-rich triarsenides (skutterudite) and diarsenides (safflorite), giving way successively to Ni-diarsenides (both rammelsbergite and pararammelsbergite), intermediate and Ni-monarsenides (maucherite and niccolite), Ni-Co- and Fe-sulpharsenides (gersdorffite, cobaltite) and Ni-Co-sulphides (millerite; bravoite; linnaeite-group

minerals including Co-rich members), followed by base metal sulphides (pyrite, marcasite, chalcocite, bornite, chalcopyrite, sphalerite, galena) and finally carbonates (calcite, siderite) and Fe-hydroxides (goethite) (Hoeve, 1981; 1984).

Uranium mineralization above the unconformity extends up to 100-150 m into the sandstone. It differs from the ore at the unconformity and in the basement in that the associated minerals are dominated by sulphides rather than arsenides, although small quantities of niccolite, gersdorffite and cobaltian rammelsbergite have been identified. Mineralization occurs as fine-grained, dispersed impregnations and as fracture-filling veins which are invariably enveloped by bleached haloes from which the hematite pigment has been removed. Base metal sulphides (galena; Cu-sulphides) are common as central vein fillings on top of euhedral quartz or as independent veins up to 10 cm wide.

The mineralization consists of a dull pitchblende which may form massive aggregates. This pitchblende is accompanied by coffinite and, in paragenetic order, pyrite plus marcasite, chalcocite, bornite, chalcopyrite, sphalerite, galena, covellite; arsenides, where present, precede the sulphides. Siderite is abundant, replacing the clay matrix as well as quartz, and carries inclusions of sulphides and goethite. Dolomite and euhedral quartz are locally present coating fracture walls and forming a substratum to the sulphides. Jordisite is probably present because much of the core develops an intensive Mo-bloom when exposed to the atmosphere. Athabasca sandstone heavy mineral bands with hematite-rutile pseudomorphs are replaced by ore minerals with preservation of texture, giving rise to chalcopyrite-rutile, siderite-rutile or coffinite-rutile pseudomorphs after ilmenite.

From these observations it becomes clear that at Midwest the temporal paragenetic succession is matched by a crude spatial zonation in which early, high-grade, arsenide-rich ore at the unconformity is surrounded by later, lower-grade, sulphide-dominated ore in the sandstone. Relationships at other deposits are essentially similar. At Collins Bay B, sulphidic, second-phase mineralization forms an envelope of lower-grade ore around the high-grade ore body at the unconformity (Jones, 1980, 1981; Jones and Heine, 1985). At McClean Lake, Wallis et al. (1984) describe arsenide- and sulphide-facies mineralization, with the sulphidic ore being external to the arsenide-rich ore.

At Midwest, as at most other deposits, the sulphidic phase of mineralization is developed in association with euhedral quartz veins and is invariably accompanied by bleaching of hematitic host rocks. At Rabbit Lake, second-stage mineralization is characterized by pyrite, marcasite, Cu-Pb-Zn-sulphides, siderite and chamosite associated with abundant euhedral quartz-carbonate-tourmaline (dravite) veins and extensive bleaching superimposed upon rocks that were altered and hematized during the first phase (Hoeve and Sibbald, 1978). At Dawn Lake, base metal sulphides are associated with euhedral quartz veins and zones of intensive, vuggy, silicification (Clark and Fogwill, 1985).

The timing of these two phases of mineralization may be inferred from published age determinations (Table 7). At the Midwest deposit, first-stage, lustrous pitchblende was dated at 1331 ± 8 Ma (U-Pb), whereas for second-stage, dull pitchblende in basement ore an age of 1094 ± 27 Ma (U-Pb) was obtained (Worden et al., 1985). Third stage sooty pitchblende was not dated, but by analogy with other deposits, it probably represents the basin-wide remobilization event at 300-250 Ma,

since an episode of lead loss was recognized at 265 ± 50 Ma (Baadsgaard et al., 1984; Cumming et al., 1984). Rb/Sr isochron dating of illitic clays at the Midwest deposits gave an age of 1337 ± 11 Ma for host rock alteration, coeval with the first stage of mineralization.

At Dawn Lake an age of 1100 Ma (U-Pb) is reported for second-stage, dull pitchblende, whereas first-stage, lustrous pitchblende was not dated (Clark and Fogwill, 1985). At Collins Bay first-stage pitchblende yields ages within the range 1238-1384 Ma (U-P) and no dates are given for second-stage pitchblende (Jones, 1981). A recent Sm-Nd dating of pitchblende renders an age of 1281 ± 80 Ma for primary mineralization (Fryer and Taylor, 1984), essentially identical to a K-Ar age of 1296 ± 34 Ma for illitic alteration clays (Tremblay, 1983). At Rabbit Lake, lustrous, first-stage pitchblende crystallized within the interval 1345-1281 Ma (U-Pb), and is followed by second-stage pitchblende, dated at 1078 ± 26 Ma (U-Pb), which is associated with base metal sulphides, euhedral quartz and dravite (Cumming and Rimsaite, 1979; Hoeve et al., 1982). Late sooty pitchblende rendered an age of 215 ± 40 Ma (U-Pb). At the McClean deposit, the initiation of illitic host rock alteration was dated at 1318 ± 3 Ma (Ar^{40}/Ar^{39}, Hall et al., in press), coeval with primary mineralization in the other deposits.

It appears from this review that first-stage, arsenide-type, mineralization was formed at ~1330-1250 Ma and second-stage, sulphide-type, mineralization at ~1100-1050 Ma. A possible third pulse of mineralization at ~900-700 Ma (Table 7) still lacks paragenetic definition.

HOST ROCK ALTERATION

Host rock alteration associated with unconformity-type uranium deposits in the Athabasca Basin affects basement rocks, the overlying sandstones and the palaeoweathering profile. It is visually manifested by bleaching/hematitization phenomena, loss of coherence of the rocks and, in the immediate vicinity of high-grade ore at the unconformity, by development of both an envelope of almost pure clay and a shell of secondary hematitization. In several deposits, a zone of drab, olive-green alteration is developed in close spatial association with the shell of secondary hematitization surrounding the high-grade ore body. Tourmalinization is prominent in some deposits, whereas in others it is lacking altogether. Host rock alteration is also accompanied by the destruction of graphite, by the formation of hydrocarbons and by extensive alteration of clay mineral species, which, though not detectible in hand specimen, is readily identified by means of XRD analysis.

The following account of relationships between the various manifestations of host rock alteration and stages of mineralization is based upon studies of the Midwest, Rabbit Lake and Collins Bay B deposits, supplemented by an interpretation of literature data for the McClean deposit (Wallis et al., 1984).

Clay Mineral Alteration

The clay mineral assemblages consist of illite, chlorite and kaolinite in varying proportions. Smectites and mixed-layer minerals are absent. Kaolinite is found in the Athabasca Group rocks and also in the upper part of the sub-Athabasca palaeoweathering profile, where it forms a laterally continuous layer immediately underlying the unconformity and approximately coinciding with the oxidized, hematitic portion of the

profile. Illite is present both above and below the unconformity, in sandstones as well as in weathered and altered basement rocks. Contrary to diagenetic the illites of unaltered sandstone which are mixtures of 1 M and 2 M polytypes, the characteristic polytype of the alteration illites is 3T. In the immediate vicinity of high-grade ore at the unconformity the alteration clay consists entirely of 3T illite, whereas farther away 1 M and 2 M polytypes appear. Chlorite is practically restricted to the basement although in severely altered rocks it is also found in the basal sandstones above the unconformity. Chlorite is predominantly of the di,trioctahedral type (sudoite), although regular trioctahedral chlorites are found in the lower part of the palaeoweathering profile and also in the innermost portions of the alteration haloe.

Of the many longitudinal and cross-sections studied for the various deposits (Hoeve, 1981; 1982; 1983) a few representative examples are shown here to illustrate characteristic relationships (Figures 8,9,10, 11,12,13). Clay mineral contours are drawn at 80%K, 60%K, 45%K, 20%K and 10%K, of which the 45% contour outlines the limit of the alteration haloe against diagenetic background values in the lower Manitou Falls Formation (Table 4). Of the four deposits studied, Key Lake shows a different type of clay mineral haloe and will for this reason be discussed separately.

Midwest, Collins Bay, McClean Lake: In these three deposits, the clay mineral contours delineate an illitization haloe which is many times the size of the actual ore bodies and which decreases in intensity with distance from the ore-bearing fault zones (Figure 8,9,11,12,13). The haloe is roughly plume-shaped above the unconformity, consisting of

a core with close to 100% (rel.) illite and successive less intensively altered shells grading outwards into unaltered sandstone.

For the Midwest deposit, a sufficient number of cross-sections have been studied that the alteration haloe can be reconstructed in three dimensions (Figure 12a,b,c). Although illite alteration is present along the full 6 km strike length of the mineralized structure, the haloe reaches its greatest dimensions in association with high-grade ore. Over the main and northern ore bodies, as well as over the inferred third hydrothermal centre in between (Hoeve, 1984), the haloe forms more or less circular, pipe-like features which rise through 200 m of sandstone to the top of the bedrock, where they are truncated by the present erosional surface. Judging from the configuration of clay mineral contours the original vertical extent may well have been in the order of 400-500 m or more. Laterally along the unconformity, the haloe displays extended wings, continuing more than 800 m outward from the ore bodies (Figures 8,12). At the northern and southern extremes of the ore-bearing structure, the illite alteration gradually dies out along strike.

At Collins Bay the ore body is covered by only 20-30 m of sandstone, so that the vertical constitution of the clay mineral alteration haloe cannot be studied. Nonetheless, observed relationships are similar to those at Midwest, with the haloe extending laterally for 800-900 m from the Collins Bay thrust fault (Figure 9).

At the McClean Lake north zone, the development of the illite alteration haloe is similar to that at Midwest, whereas over the south zone it is highly irregular (Figure 10). Here, areas of intensive illitization immediately above the unconformity and near the surface are

separated by an intervening zone of unaltered or weakly altered kaolinitic sandstone.

For all three deposits, host rock alteration in the metamorphic basement is restricted to the immediate vicinity of the ore-bearing fault zone. Where the fault zone intersects the unconformity, the lateral continuation of the kaolinite layer at the top of the palaeo-weathering profile is interrupted owing to conversion of kaolinite into illite and chlorite (sudoite). Extending downwards as a narrow funnel straddling the ore zone, basement rocks within the haloe are intensively altered to an illite/chlorite assemblage, locally up to 150 m below the unconformity.

Esquevin diagrams for altered and unaltered sandstones are generally similar (Figure 14) and characteristic of Dunoyer de Segonzac's field III (Figure 7), indicating that the alteration illite crystallized under essentially the same conditions of high-grade diagenesis and deep burial as the diagenetic assemblages in adjacent unaltered rocks. In detail, however, the diagrams differ in that the clusters of data points for altered sandstones are shifted towards higher values of the I (002/001) ratio and towards slightly lower values of the Kübler index which may reflect somewhat higher temperatures at the site of the deposits. As such a shift is consistently observed in all deposits studied to-date (Table 9), this suggests that higher (002)/(001) peak intensity ratio of the alteration illites may constitute an additional criterion to recognize alteration assemblages.

Key Lake: At the Key Lake deposit, host rock alteration below the unconformity is similar to that in the other deposits, whereas in the sandstones it is markedly different (Figure 11). As in the other depos-

Table 9. Illite peak intensity ratios I(002)/I(001) for altered and unaltered sandstones.

	MFa+b regional	Unaltered	Altered	Illitic sandstone, Key Lake
Regional	0.35±0.12			
Midwest		0.34±0.24	0.47±0.57	
McClean		0.35±0.19	0.45±0.12	
Collins Bay			0.57±0.12	
Key Lake			0.52±0.11	0.49±0.10

its, the kaolinite-rich top zone of the palaeoweathering profile is interrupted in the vicinity of the ore body owing to conversion of kaolinite into illite plus chlorite. Chlorite in the top part of the profile and in the upper portion of the ore body is of the Mg-Al-rich, di,trioctahedral type (sudoite), whereas farther down it gradually gives way to trioctahedral types. Host rock alteration extends along faults to at least 150 m below the unconformity.

Above the unconformity, the clay mineral contours delineate a haloe of intensive kaolinitization ($\geq$90% K rel.) which is emplaced in an anomalously illitic sandstone ($\geq$ 80-90% I rel.). Locally, along fractures, the kaolinite haloe protrudes somewhat below the unconformity and appears to be superimposed upon altered basement. Other drill holes in the general area show late, fracture-controlled kaolinitization similar to that observed in the regional studies of clay mineral distribution in the Athabasca Group.

The Esquevin plot for the illites of the anomalously illitic sandstone (Figure 14d) shows that they share the high (002)/(001) peak intensity ratios of the alteration illites of other deposits.

Taken together, the evidence indicates that two different clay mineral haloes are present at Key Lake: (a) an illite (± sudoite) haloe, probably associated, as in other deposits, with the first phase of mineralization, formed under conditions of deep burial and high-grade diagenesis; (b) a superimposed kaolinite haloe, whose formation may be related to remobilization of the sandstone portion of the ore body at approximately 300-250 Ma and which may correlate with the late-diagenetic, fault-controlled kaolinitization referred to previously.

Bleaching

Visually, the most conspicuous aspect of host rock alteration is the chemical reduction of ferric iron as manifested by striking white and creamy, locally olive-green, colour effects due to bleaching of the normally purple or pink sandstones of the lower Manitou Falls Formation. The sub-Athabasca palaeoweathering profile is similarly bleached where affected by host rock alteration. The bleaching is fracture and permeability controlled, forming haloes around micro-fractures, joints and faults, and laterally advances along zones parallel to lithological bedding. Frequently, the bleached rock is separated from the purple hematitic rock by a narrow zone of orange-red to brick-red coloration, which seems to constitute the first effects of chemical reduction of ferric iron and may arise from a decrease in grain size of the specular hematite of the purple rock. These relationships are observed on a millimetre as well as a metre-scale.

Bleaching is in part accompanied by sulphide-stage mineralization, represented by fracture fillings or massive replacements of the interstitial sandstone matrix by siderite and base metal sulphides. In rocks that are rich in hematite, bleaching may merge into drab olive-green alteration (see below).

Incipient bleaching starts with the development of isolated bleached spots that are dispersed in an otherwise hematitic sandstone. Increased bleaching gives the rock a mottled appearance and ultimately renders a pervasive white or creamy colour. Concomittently, black detrital heavy mineral bands become less and less conspicuous, as the hematite-rutile pseudomorphous grains are replaced by grey or brownish siderite-rutile pseudomorphs.

Systematic core logging at Midwest revealed that the ore-bearing zone is surrounded by a distinct haloe of intensively bleached rock which reaches many times the size of the actual ore bodies. In cross-section the haloe is roughly plume-shaped above the unconformity and funnel-shaped below it in the basement. In the basal sandstone, bleaching extends laterally for approximately 100 m on either side of the ore-bearing zone. The vertical magnitude of the haloe is variable and seems to be roughly proportional to the intensity of mineralization, as it is only directly over major ore accumulations such as the northern and main ore bodies that bleaching extends to the top of the bedrock, where it is truncated by the present surface.

In basement rocks the haloe has its greatest dimensions at the unconformity and narrows down rapidly to a funnel shape that straddles the fault zone. In plan view the extent of bleaching in basement rocks at the unconformity (Figure 12d), largely coincides with the trace of the graphitic conductor. At higher levels above the unconformity, the coherent bleaching haloes resolve into more or less circular, pipe-like features, associated with distinct ore accumulations.

Although the bleaching and clay mineral haloes are largely congruent, in detail they show cross-cutting relationships (Figure 13). This is dramatically illustrated in Section #6890 of the Midwest deposit, where clay altered rocks are not bleached (drill holes #158,159) and unaffected rocks are pervasively bleached (drill-hole #153).

These relationships, together with the previously noted association of bleaching with euhedral quartz veins and sulphidic mineralization, allow the formation of the bleaching haloe to be correlated with the second phase of mineralization at 1100-1050 Ma.

Table 10. Partial chemical analyses of clay envelope compared to unaltered sandstone; 1-10: hematite-rich clay from zone of secondary hematite; 11-20: white, bleached clay; 21: unaltered sandstone (MFa+b, average of 150 samples).

	Al_2O_3	TiO_2	Fe_2O_3	CaO	MgO	K_2O	Na_2O
1	25.8	.80	26.0	.14	.62	6.55	.31
2	25.9	1.85	20.2	.16	.99	5.70	.48
3	20.5	1.00	3.14	.05	.30	1.43	.25
4	16.9	.39	48.7	.10	.44	3.66	.57
5	23.0	1.98	36.4	.12	.58	5.46	.52
6	21.4	1.31	48.3	.30	.73	5.02	.51
7	33.4	.79	4.09	.14	4.71	6.16	.73
8	19.0	.84	39.8	.22	.87	2.70	.54
9	19.3	.74	43.7	.09	2.92	2.35	.80
10	30.1	.53	20.2	.19	.85	8.08	.58
11	28.4	1.11	2.49	.20	8.36	1.68	.59
12	21.1	1.02	2.51	.23	6.18	6.67	.62
13	34.1	2.97	1.11	.06	5.09	5.14	.62
14	30.5	5.90	1.13	.14	.74	7.90	.47
15	35.1	.74	.45	.09	1.09	9.82	.58
16	28.1	.81	1.96	.15	.83	8.96	.51
17	31.8	.25	.73	.15	1.07	9.42	.57
18	18.8	.27	1.60	.02	1.09	2.20	.25
19	33.9	.54	.89	.05	.70	9.29	.47
20	33.6	2.26	.60	.07	.64	9.46	.43
21	1.92	.09	1.35	.03	.03	.13	.12

Table 11. Partial chemical analyses of sandstones affected by secondary hematitization compared to unaltered sandstone; 1-7: hematite-rich altered sandstone samples from zone of secondary hematite; 8: unaltered sandstone (MFa+b, average of 150 samples).

	Al_2O_3	TiO	Fe_2O_3	CaO	MgO	K_2O	Na_2O
1	6.5	.65	7.27	.03	.29	1.23	.08
2	7.0	.34	6.09	.04	.26	1.00	.11
3	4.3	.25	2.37	.03	.10	.74	.07
4	3.7	.15	7.39	.09	.10	.14	.07
5	5.1	.16	3.86	.03	.08	.97	.20
6	3.0	.24	20.40	.05	.03	.25	.06
7	2.2	.16	21.36	.62	.71	.34	.12
8	1.9	.09	1.35	.03	.03	.13	.12

Loss of Coherence; Clay Envelope; Quartz Dissolution

The second conspicuous effect of host rock alteration is a loss of coherence of the affected rocks, although this feature is not nearly as visually striking as is the bleaching. The sandstone above the ore bodies is generally fractured or brecciated and contains open veins of euhedral quartz, which may also carry sulphide-stage mineralization. As the ore bodies are approached, the friability of the sandstone increases progressively until the rock has all but entirely disintegrated. Concomitantly, the clay content of the rock increases and the shape of the quartz grains becomes more angular; fracture-controlled clay seams begin to appear and ultimately high-grade ore at the unconformity is found embedded in what appears to be a clay envelope up to several metres thick, in which a few angular quartz grains may still remain. Similarly, below the unconformity the clayey altered basement underlying high-grade ore at the unconformity is marked by a distinct lack of quartz.

The increase in clay content towards the ore bodies is illustrated by the contoured cross-section of the Midwest deposit (Figure 15), showing the weight percentage of the $\leq 2\mu m$ clay-size fraction of the rock. Near the high-grade ore bodies, the host rocks become enriched by a factor of 5-10 times the percentage of the $\leq 2\mu m$ fraction in the sandstone farther away.

Petrographic examination of altered sandstones reveals textural evidence indicative of desilicification, i.e., disappearance of quartz overgrowths (Figure 15), irregular dissolution embayments, development of angular shapes, presence of unsupported, corroded grains isolated in a clay matrix (plate 1F).

The loss of coherence is interpreted as to result from corrosion of quartz cement, and the development of a distinct clay envelope as to represent residual clay enrichment resulting from quartz dissolution. Should this be the case, then the development of the clay envelope must have been accompanied by a considerable volume loss.

In the sandstone overlying the Midwest deposit, the boundary between an upper sandy unit and a lower conglomeratic unit provides a suitable reference horizon against which possible volume effects can be evaluated. An analysis by Sibbald (written comm. 1984) indicates that the level of the reference horizon indeed does show a dip over high-grade ore.

An additional argument in favour of a residual enrichment origin for the clay envelope derives from the observation that the clay-rich rocks tend to be enriched in relicts of sandstone heavy mineral bands. These rocks contain rounded detrital grains of zircon, rutile and relicts of the earlier described hematite-rutile pseudomorphs after ilmenite, which are embedded in a clay matrix and enclosed by coffinite and arsenides. The hematite component of the pseudomorphs has been replaced by coffinite, arsenides and clay minerals; however, the rutile lamellae still follow rhombohedral directions of the original ilmenite.

The above relationships point to a direct connection between first-stage mineralization, host rock disintegration and clay development. The evidence indicates that the formation of the high-grade ore body at the unconformity was accompanied by massive quartz dissolution leading to residual enrichment of clay and collapse of the overlying sandstone, concomitantly with the illitization of sandstone and basement rocks.

Secondary Hematitization

At the Collins Bay B and Midwest deposits, the high-grade ore at the unconformity is surrounded by a zone of intensive secondary hematite. This secondary hematite forms a fine-grained, powdery pigment of a reddish-brown to chocolate-brown colour which occurs as coatings around quartz grains, as pore fillings and in extreme cases as replacements of interstitial clay. In appearance and mode of occurrence it is distinct from and should not be confused with, either the purple, diagenetic, coarser grained specularite, or the finely divided, brick-red hematite marking the first effects of bleaching. Chemical analyses illustrate that the zone of secondary hematitization is strongly enriched in iron (Tables 10, 11).

Despite partial destruction by large-scale superimposed bleaching, the overall configuration of the zone of hematitization can be readily reconstructed from its preserved remnants (Figure 8). The zone forms a shell which surrounds the high-grade ore body, although there is a partial overlap between hematitization and ore. The shell partially coincides with the clay envelope, but the outer boundary of hematitization extends further from the high-grade ore. Internally to the shell, the rock is pervasively reduced (grey colours prevail) and intensively mineralized, whereas externally, relicts of diagenetic specular hematite are present that have escaped the superimposed, large-scale bleaching.

Based on these relationships, we interpret the shell of secondary hematitization as marking the position of a distinct redox front related to the formation of the high-grade ore body during the first stage of mineralization. Inside the front ore constituents were precipitated; outside of the front diagenetic specular hematite was initially pre-

served despite intensive clay alteration, until affected by the superimposed bleaching associated with the second stage of mineralization.

For the McClean Lake deposits, Wallis et al. (1984) describe a similar zone of hematite enrichment, which they refer to as "hematite facies" mineralization, located externally to high-grade ore. They interpret the "hematite facies" as a late feature that encroaches upon earlier arsenide/sulphide mineralization. Although we have not examined drill core from the McClean deposit, we view this interpretation with suspicion, because cross-cutting relationships at the Midwest and Collins Bay B deposits unequivocally demonstrate that in these deposits the shell of secondary hematitization is an early feature that pre-dates the large-scale bleaching, the olive-green alteration (see next section) and the second, sulphide-stage of mineralization. We consider it significant that at the Midwest and Collins Bay B deposits, and apparently also at the McClean Lake deposits (Wallis et al., 1984, pp. 89; figure 7.14), the upper limit of second stage, sulphidic mineralization is external to the shell of secondary hematitization (Figure 8).

Drab, olive-green alteration

At the Collins Bay B deposit, drab, olive-green alteration forms an incoherent zone around high-grade ore which approximately coincides with and is superimposed upon the shell of secondary hematitization. It is accompanied by sulphide-phase mineralization and characterized by the ferrous-iron-rich chlorites chamosite and berthierine accompanied by illite (for nomenclature cf. Bailey, 1981; chamosite = "14Å chamosite"; berthierine = "7Å chamosite" = "septechlorite"). The olive-green alteration shows identical relationships to its host rocks as the bleaching described above. In fact, zones of bleaching and olive-green alteration

are seen to merge into one another, the green colour appearing where the host rock is particularly rich in hematite. Individual fractures filled with siderite or sulphide-phase mineralization may display zoned haloes of olive-green alteration and bleaching, the latter forming an external zone between unaffected hematitic sandstone and an innermost zone of olive-green chamositic alteration adjacent to the fracture filling. Similar relationships are noted in basement rocks at Rabbit Lake, where euhedral quartz veins of second-stage sulphidic of mineralization are associated with extensive bleaching ("pale green alteration"; Hoeve and Sibbald, 1978) superimposed upon earlier hematitic alteration ("red alteration"). In the paragenetic sequence for these quartz veins, siderite is accompanied by base metal sulphides, which occupy vein centres and vugs.

At the Midwest deposit, olive-green alteration is not well developed; where present, however, its relationships to other aspects of host rock alteration and stages of mineralization are identical to that described above for the Collins Bay B deposit.

For the McClean Lake deposits, Wallis et al. (1984) report a green, ferrous-iron chlorite alteration, which judging from their descriptions and figures seems to have similar relationships to sulphidic mineralization and bleaching as described above for the other deposit (Wallis et al., 1984; pp. 91-93; figure 7.14; "Rabbit ears" Figure 7.15).

On the basis of our observations we interpret the olive-green alteration as a manifestation of the large-scale, superimposed bleaching that accompanied the second, sulphidic stage of mineralization. The development of the olive-green colour is a consequence of the formation of ferrous-iron chlorites in hematite-rich rocks.

Tourmalinization

Tourmalinization is developed in some deposits only. At the Collins Bay B deposit (Jones, 1980) it is not present at all, whereas at the McClean Lake and Midwest deposits, boron geochemical haloes, but no significant tourmalinization, have been recognized (Sopuck, 1981; Wallis et al., 1984). By contrast, at the Rabbit Lake, Key Lake and Cluff Lake deposits and at various prospects along the southern and eastern margin of the Athabasca basin, tourmalinization is prominent. In these deposits tourmalinization is manifested by fine-grained, radial-textured, spherulitic aggregates of colourless dravite,which may occur dispersed through the altered rocks or form distinct vein fillings that are overgrown by euhedral quartz. Dravite also occurs as colourless, oriented overgrowths around variably coloured detrital grains in the sandstone and around metamorphic grains in altered basement rocks.

Although it may be difficult to assign the tourmalinization a distinct position in the mineralization and alteration sequence, the association of dravite veins with euhedral quartz suggests a link with the second stage of mineralization. However, tourmalinization may have started during the first stage, as indicated by paragenetic relationships at the Rabbit Lake deposit (Hoeve, 1977; Hoeve and Sibbald, 1978).

Graphite destruction; Hydrocarbon formation

Destruction of graphite and formation of solid hydrocarbon beads in altered rocks was first reported at Rabbit Lake, (Hoeve and Sibbald, 1978) and have since been recognized in most deposits of the basin. At Rabbit Lake, graphite-bearing rocks are absent from the thoroughly altered upper portions of the deposit; however, mining activities have since exposed graphite-bearing dolomitic marbles and graphite-rich

pelitic rocks at lower levels of the deposit. Intermediate levels are characterized by partially altered graphitic rocks, which contain an abundance of solid, hydrocarbon "buttons". At these levels, relict blocks of partially altered graphitic rocks carrying hydrocarbon may be seen isolated in thoroughly altered, brecciated and chloritized rocks. Fractures in graphitic rocks are flanked by bleached, pale green, chloritized, graphite-free haloes. Similarly, veinlets of early (Stage 1) cherty quartz are surrounded by chloritized envelopes devoid of graphite. In thin section, graphite flakes in partially altered rocks, contrary to those in unaltered rocks, display frayed margins, suggestive of corrosion. Occasionally, the front of bleached graphite-free rock against non-bleached graphite-bearing rock is enhanced by accumulations of solid hydrocarbon buttons.

Similar relationships are noted in other deposits. For Collins Bay B, Jones (1980) reported that mineralized, altered graphitic metapelites contain less graphite than their unaltered counterparts. For the McLean Lake deposits, Wallis et al. (1984) noted the destruction of graphite in altered metapelites underlying high-grade ore bodies at the unconformity. For the Midwest deposit, the boundary indicating the upper limit of graphite shows a distinct depression underneath the ore body (Figure 8).

At Key Lake and Cluff Lake, high-grade ore is locally extremely rich in solid hydrocarbon matter even to the extent that it may form a cement to botryoidal and dendritic aggregates of pitchblende and Ni-Co-arsenides. Despite objections raised by organic geochemists, who seem to advocate a biogenic, algal origin for the hydrocarbon (Rouzaud, 1979; Rouzaud et al., 1981), we contend that the geological relation-

ships point to a direct connection between host rock alteration, graphite destruction, hydrocarbon formation and mineralization, that is, to a hydrothermal origin of the hydrocarbon matter. Landais and Dereppe (1985) reach this same conclusion for the Cluff Lake deposits.

At Rabbit Lake, cross-cutting relationships between early veins of cherty quartz (stage 1) flanked by bleached zones containing hydrocarbon, and euhedral quartz veins (stage 2) filled with calcite and hydrocarbon buttons, suggest that graphite consumption and hydrocarbon formation took place during both the first and the second stage of mineralization.

BASINAL HISTORY, DIAGENESIS, ALTERATION AND MINERALIZATION

The remarkable coherence in the sequence of events documented in the deposits implies that the stages of mineralization and host rock alteration do not reflect local peculiarities of single ore forming systems, but rather are manifestations of basin-wide events. Published dates, obtained by various radiometric methods applied to alteration clays and unaltered Athabasca Group rocks (table 1) suggest that the overall episode of mineralization and alteration was coeval, not only with the period of diabase intrusive activity, but also with illite diagenesis in the unaltered sandstones away from mineralization.

The first stage of mineralization took place during the 1350-1250 Ma interval, 100-150 Ma after deposition of the Athabasca Group (Figure 16). This stage represents the main ore-forming event which led to the formation of the high-grade primary ore bodies, characterized as described above by an assemblage of lustrous pitchblende, Ni-Co-arsenides and -sulpharsenides. Host rock alteration is marked by clay alteration (illite ± chlorite), quartz dissolution, residual clay enrichment, collapse of the sandstone, the formation of a shell of secondary hematite, destruction of graphite and the formation of hydrocarbon (table 12). Clay mineral characteristics indicate that these processes took place during an advanced stage of basin evolution under conditions of deep burial and high-grade diagenesis with temperatures in the order of 150-200°C.

The second stage of mineralization, which took place 150-200 Ma after the first, during the 1100-1050 Ma interval, is characterized by a dull reflecting pitchblende accompanied by base metal sulphides, siderite and ferrous iron chlorites. Host rock alteration is represented by

	Mineralization	Alteration
Stage 3: 300-250 Ma	sooty pitchblende/coffinite also pitchblende/hematite	superimposed kaolinitization
Stage 2: 1100-1050 Ma	low-grade envelope dull pitchblende base metal sulphides siderite	extensive bleaching euhedral quartz veins, tourmaline, siderite, calcite dolomite, goethite, Fe^{2+}-chlorite graphite destruction solid hydrocarbon beads.
Stage 1: 1350-1250 Ma	high-grade core lustrous pitchblende Ni-Co-arsenides U/Ni-zonation Arsenide zonation	illitization/chloritization quartz dissolution; dolomite dissolution residual clay enrichment: clay cap secondary hematite (redox front) graphite destruction solid hydrocarbon beads

Table 12: Stages of uranium mineralization and remobilization with associated host rock alteration, recognized in unconformity-type deposits of the Athabasca basin.

extensive bleaching superimposed upon first-stage alterations, by formation of veins of euhedral quartz and tourmaline, the development of drab, olive-green alteration and probably by renewed graphite destruction and hydrocarbon formation. Fluid inclusion studies of euhedral quartz (Pagel, 1975a,b) indicate continuing high-grade diagenetic conditions, at temperatures in the 180°-220°C range. A clustering of U/Pb dates at approximately 900-700 Ma (Table 7) points to a possible additional event, which has not been paragenetically defined and remains to be confirmed.

The third significant event in the history of the deposits is represented by the 300-250 Ma episode of remobilization and accompanying kaolinitization which are superimposed upon the earlier alterations.

The above relationships indicate an intimate connection between ore formation, diagenesis and basinal history. Whereas the first and second stages of mineralization and alteration are related to prograde burial and diagenesis, the third stage of remobilization and kaolinitization is inferred to be connected to uplift and retrograde diagenesis.

The Athabasca basin constitutes a tectonically active sedimentary basin whose development is broadly similar to that of other intracratonic basins. Two fundamentally different stages are recognized in the evolution of many sedimentary basins: an early tectonic, rifting phase and an ensuing non-tectonic, thermal relaxation phase, although the history of a basin may be complicated by episodic tectonic reactivation (McKenzie, 1978; Bally, 1982; Watts et al., 1982; Kingston et al., 1983; Jarvis, 1984; Perrodon and Masse, 1984; Sleep, 1984). The early tectonic phase is marked by rapid subsidence and high rates of sedimentation and typically lasts only a few to some ten million years. The

post-tectonic phase is characterized by the development of a broad platform-type basin, as a result of areal flexuring in response to sediment loading and thermal contraction over a prolonged period lasting in the order of 50-100 Ma or more.

From the stratigraphic and diagenetic development as interpreted in this study, several stages of contrasting structural characteristics can be distinguished in the Athabasca basin which correspond to the evolutionary stages of other sedimentary basins (Figure 16). Radiometric evidence (Bell, 1981) suggests that during the interval 1550-1500 Ma, immediately preceeding the formation of the basin, the region had undergone mild uplift of crustal segments, possibly accompanied by minor magmatic remobilization. The actual basin formation was initiated by an early tectonic phase, characterized by a tensional regime, which generated the Cree, Mirror and Jackfish sub-basins. This is consistent with Walcott's (1968) interpretation of a gravity study suggesting that the basin is underlain by northeasterly trending twin graben structures. The early stratigraphic development was controlled by the three sub-basins as they were filled in with the lower Manitou Falls (MFa + b), Lazenby Lake (LzL) and Fair Point (FP) formations. The subsequent advance of the upper Manitou Falls (MFc + d) formation over almost the entire basin led to coalescence of the sub-basins, marking the transition to the ensuing non-tectonic phase. Towards the end of this period the sedimentary cover was approximately 1000 m thick (Figures 4,5,6,16). Burial compaction and diagenetic recrystallization may have begun during this phase although no record is preserved in the clay mineral assemblages. Quartz overgrowths within the sandstones of the Lazenby Lake

formation, however, formed during early diagenesis and shallow burial, before significant compaction had taken place (plate 1B).

The subsequent non-tectonic phase is marked by a uniform, basin-wide stratigraphic development reflecting a general downwarp of the area and a considerable broadening of the basin. The influence of the sub-basins is no longer apparent. Grain size analysis revealed that during sedimentation of the Wolverine Point, Locker Lake and Otherside Formations, in contrast to deposition of the Manitou Falls formation, the basin margin was situated well east of its present position (Ramaekers and Dunn, 1977). Depth of burial during this phase is unknown, but is likely to have exceeded the presently preserved stratigraphic thickness of approximately 2 km (this estimate includes the Douglas and Carswell Formations preserved within the Carswell structure; Hoeve et al., 1985).

The non-tectonic phase was abruptly terminated by the intrusion of diabase dikes starting at approximately 1350 Ma which marks the onset of a phase of renewed crustal instability and tectonic reactivation of the basin. The dikes follow NW-SE trending tensional directions associated with sinistral wrench movements on the Hudsonian faults underlying the basin. Diabase intrusive activity took place over a 350 Ma period of basin development between approximate 1350 and 1000 Ma; however, uniform magmatic activity over such a long time span is unlikely, and the intrusion of the dikes may actually have been episodic in nature, even though currently available radiometric dates do not allow distinct magmatic pulses to be resolved. Coeval uranium mineralization was definitely episodic, with two distinct stages of hydrothermal activity at 1350-1250 Ma and 1100-1050 Ma, respectively, and a possible third around 900-700 Ma. We speculate that these stages may be manifestations of thermal

pulses that correlate with peaks of diabase magmatic activity and, by implication, with episodic movement on the Hudsonian faults. Some of the dikes are offset by these faults, attesting to post-intrusive tectonic activity.

By analogy with other sedimentary basins, episodes of diabase magmatism may represent periods of active deepening of the basin. For example, in the Sverdrup basin in the Canadian Arctic, Balkwill (1978) documented that episodes of mafic activity approximately coincide with periods of accellerated subsidence and increased rates of sedimentation. We suggest that the Athabasca basin reached its deepest level of burial and peak diagenetic conditions during the phase of tectonic reactivation and diabase intrusive activity. The total sediment thickness as inferred from fluid inclusion evidence was in the order of 4-5 km (Pagel, 1975).

The relatively low degree of diagenetic recrystallization in the upper Wolverine Point Formation (WPb) as compared to underlying sandstones of the Manitou Falls (MF), Lazenby Lake (LzL) and Fair Point (FP) Formations (Figure 7) is interpreted as indicative of low permeability for this clay-silt-rich unit during diagenesis. This suggests that the upper Wolverine Point Formation (WPb) may have formed a relatively impermeable seal, promoting the possible development of a geopressured zone in the underlying formations.

Following the conclusion of the second stage of mineralization at approximately 1050 Ma, and the termination of the phase of tectonic reactivation, no distinct major events are recognized in the Athabasca basin until the episode of uranium remobilization and kaolinitization at 300-250 Ma.

In a geotectonic perspective, the initiation of the Athabasca basin at 1450±30 Ma may be a distant manifestation of the Elsonian thermal event (Stockwell, 1964) which is expressed south of the Canadian Shield as a SW-NE trending belt characterized by regional metamorphism and widespread igneous activity. Though truncated by the Grenville Orogen, this zone extends from the Colorado Front Range northeastwards into eastern Canada. The well-known Silver Plume and Sherman Granite Batholiths in Colorado and Wyoming, dated at 1450-1390 Ma (Peterman and Hedge, 1968; Hills and Armstrong, 1974), were intruded during this period. Major Hudsonian faults, such as the Needle Falls Shear Zone (Figure 1), which controlled the formation of the Athabasca basin, can be traced for over 2000 km through the Dakotas and Wyoming into the Colorado Front Range (Camfield and Gough, 1977; Handa and Camfield, 1983).

In the Canadian Shield the 1350-800 Ma interval is punctuated by the intrusion of several diabase dike swarms representing episodic crustal stretching, and by intense deformation associated with the Grenville Orogeny. In the Athabasca basin, the initiation of diabase magmatism at 1350 Ma and the coeval first phase of mineralization at 1350-1250 Ma may be correlated with the intrusion of the McKenzie diabase dike swarm, dated 1300-1200 Ma (Baragar, 1977). This swarm extends through most of the Canadian Shield from Ontario to Manitoba, Saskatchewan and into the Northwest Territories, where the Coppermine Plateau basalts are considered their extrusive equivalents. The second phase of mineralization in the Athabasca basin at 1100-1050 Ma may represent a second pulse of magmatic activity associated with the same dike swarm. It appears to be coeval with the intrusion of the Keweena-

wan basalts in the Superior region, dated 1150-1000 Ma, which represents the onset of rifting prior to the Grenville Orogeny at 1000 Ma (Baragar, 1977). The poorly-defined and yet-to-be-confirmed, possible third phase of mineralization at 900-700 Ma would be coeval with the episode of post-Grenvillian rifting that accompanied or preceded the late-Proterozoic break-up of the Middle-Proterozoic supercontinent (Sawkins, 1976; Stewart, 1972, 1976; Piper, 1982; Devlin et al, 1984; Bond et al, 1985). According to Young (1984), South-Australia and the Cordilleran region of Western Canada may represent fragments of such a supercontinent which had been juxtaposed in Proterozoic times. In this context, it is noteworthy that a distinct 900 Ma mineralization or remobilization event is recognized in the unconformity-type uranium deposits of Australia (Hills and Richards, 1976; Ferguson et al., 1980; Page et al., 1980; Ferguson et al., 1983).

The 300-250 Ma episode of remobilization and kaolinitization is interpreted as a period of retrograde diagenesis related to uplift and erosion of the basin, when surface waters regained access to the deep aquifer. It was approximately during this episode that the Canadian Shield re-emerged above sea level following the Cambrian transgression (cf. palaeogeographic maps in Dott and Batten, 1971).

Should these rather speculative correlations hold true, it would appear that the Athabasca basin contains a record of the major crustal stretching events that affected the North-American continent suggesting that a similar record may be contained in other intracratonic basins that are underlain by stable crust.

METALLOGENESIS: SANDSTONE-BASEMENT INTERACTION DRIVEN BY CONVECTION

The diagenetic-hydrothermal metallogenetic model, proposed earlier as a working hypothesis (Hoeve and Sibbald, 1976; 1978; Hoeve et al., 1980), attributes the origin of unconformity-type uranium deposits to fluid interaction between oxidized, basal Athabasca Group sandstones and the underlying reduced basement, during an advanced stage of basin evolution. Since the basin floor consists of high-grade metamorphic rocks of inherently low permeability, all such interaction must have been restricted to the immediate vicinity of the unconformity, thus accounting for the observed spatial relationships between ore and the palaeoweathering profile. A basin-wide redox boundary is located at the base of the hematitic, upper portion of the palaeoweathering profile, a few metres below the unconformity.

The model envisages that oxidized Athabasca pore waters penetrated below the unconformity via fault zones, reacted with graphitic basement rocks and re-appeared above the unconformity as reduced fluids containing hydrocarbons and carbon dioxide. At discharge areas of diagenetic-hydrothermal flow systems, mineralization resulted from interaction of the newly formed reduced fluids emerging from the basement with metal-bearing, oxidized pore fluids of the basal Athabasca aquifer. Mineralization, therefore, was subjected to hydrodynamic control. Exceptionally high grades as encountered in many unconformity-type deposits may have developed in a situation of prolonged stationary hydrodynamic conditions, in a system marked by continuous replenishment of ore constituents and continuous generation of hydrocarbon reductants. Repeated oscillations between oxidizing and reducing conditions as documented at Rabbit Lake (Hoeve, 1978; Hoeve and Sibbald, 1978) may have arisen from

changes in the hydrodynamic regime in response to modification of the plumbing system by intermittent brecciation.

The results of the present study add strong support to this concept of diagenetic-hydrothermal interaction between oxidized sandstone and reduced basement. The configuration of clay mineral contours and the geometry of the bleached haloes indicate that at the site of mineralization, reduced fluids emanating from basement faults were discharged into the basal Athabasca sandstone where their advance was fracture- and permeability-controlled. The importance of permeability may possibly be illustrated by the differential development of the illitization haloes over the north and south zones of the McClean deposits (Figure 10).

Although fluids evidently emerged along a considerable strike length of the mineral-bearing structures, major discharge must have been limited to distinct centres, because it is over discrete ore bodies that the alteration haloes form plumes that rise for hundreds of metres into the overlying sandstones (Figures 8,9,10,12). This situation is analogous to that of a series of springs controlled by a fault lineament at the earth's surface, where groundwater may seep out all along the lineament, but significant discharge is restricted to springs.

Comparing the configurations of the clay mineral and bleached haloes, it is clear that the reduced fluids associated with the first and second stage of mineralization issued from the same discharge centres but under different hydrodynamic regimes (Figure 12). During the first stage, fluid interaction was such that a stationary redox front was established, as outlined by the shell of secondary hematite (Figure 8). Further characteristics, including the symmetrical development of the clay alteration pipes and their location over high-grade ore, imply

an absence of significant cross flow in the Athabasca aquifer at the deposit sites. During the second stage, reductants were released from the basement on a large scale as attested to by the wide-spread superimposed bleaching. No evidence of a stationary redox interface is found and the lateral asymmetry of the bleached haloe at the Midwest deposit, for example, indicates a certain cross-flow in the basal Athabasca aquifer.

The configurations of the haloe of quartz dissolution and residual clay enrichment (Figure 15) imply that the fluids emerging from the basement were undersaturated with respect to quartz. As the solubility of quartz depends almost exclusively on temperature, the discharging fluids were probably warmer than the ambient temperatures in the basal sandstone. Within the 150-200°C range, a temperature increase of 30-50°C raises quartz solubility by a factor of approximately 2, from 50-75 ppm to 100-150 ppm (Siever, 1962). Assuming more or less uniform fluid flow, the configuration of contours outlining the disappearance of quartz overgrowths (Figure 15) probably portrays the configuration of palaeo-isotherms at the sites of diagenetic-hydrothermal discharge and, consequently provides information on the thermal structure of the ore-forming systems.

The source of ore metals in the unconformity-type deposits still is a contentious issue. Although chemical contributions by both sandstone and basement form an inherent aspect of our concept of sandstone-basement interaction, we believe that the basement played a subordinate role as a supplier of ore constituents. The configuration of the hematitic redox front indicates a sandstone source for the main ore constituents uranium and arsenic, which are transported in the oxidized form. It is

difficult to see how these constituents could have been transported by reducing fluids emerging from the basement and be precipitated the moment these fluids were discharged into the oxidizing medium presented by the basal red bed sandstones. Further arguments in support of a sandstone source derive from the realization that the element assemblage of the deposits, comprising U, Ni, Co, Mo, Fe, V, Cu, Pb, Zn, Ag, Au, As, Se and S (some of the Cluff Lake deposits also contain Bi and Te) appears to be characteristic of the red bed environment. The red-bed associated sandstone-type uranium deposits of the western United States carry a closely similar element suite (Shoemaker, 1959; Finch, 1970; Harshman, 1974), as do red-bed associated, sandstone copper deposits and copper shales, such as those of central and east Africa and Europe (Rose, 1976; Annels, 1982; Knutson and Donnelly, 1983; Annels and Simmonds, 1984; Brown, 1984). Occurrences of U-Ni-Co mineralization in the Katanga system of east Africa are spatially and stratigraphically associated with the stratabound Zambian and Shaban copper deposits of the Copper Belt (Ngongo-Kashisha, 1975; Meneghel, 1981). The same suite is strikingly present in reduction spots containig dark cores of ore minerals ("fish eyes") which are found in many red-bed sequences. Cupriferous reduction spots in Mississippian red-beds of New Brunswick (Canada) contain Ag, Hg and V in addition to Cu (Van de Poll and Sutherland, 1976). In uraniferous reduction spheroids in Mississippian red-bed sandstones of Nova Scotia, U is accompanied by Cu, Ni, As and Hg (Dyck and McCorkell, 1981). Spectacular examples of zoned reduction spots rich in U, V, Ni, Co, Cu, Ag, As are reported for Permo-Triassic red beds in southwest England (Harrison, 1975; Durrance et al., 1978; 1980).

We have previously argued (Hoeve and Sibbald, 1978; Hoeve et al., 1980) that the ore metals of unconformity-type deposits may have been contained in detrital ferromagnesian minerals and feldspars from which they were liberated in the course of post-depositional alteration. However, in view of our present findings that detrital feldspar, biotite and mica were scarce in the Athabasca Group rocks, different carriers must be considered. Ferric iron hydroxides and kaolinitic clay, which were the main detrital components other than quartz in the original sediments of the fluviatile sequences, may offer a realistic alternative. Experimental work has shown that both may act as scavengers for a variety of metals (Grimme, 1968; Jenne, 1968; Van der Weyden and Langmuir, 1976; Koons et al., 1980; Ames et al., 1983). In first-cycle red beds, leachable U, Ni, Co, Mo, Cu, Pb, Zn are primarily associated with ferric iron oxides, particularly with amorphous and poorly crystalline hydrated forms, and to a lesser degree with clays (Zielinski et al., 1983). In lateritic weathering profiles, as preserved in the Athabasca basin floor, U, Ni, Co, V, Mo, Cu, Pb, Zn, As, Se, B are intimately associated with ferric iron hydroxides (Wedepohl, 1978; Mosser et al., 1985). In modern rivers such as the Amazon and the Yukon, iron and manganese hydroxides represent the major transport mechanisms for Ni, Co and Cu (Gibbs, 1977).

In view of these considerations it seems to us that the main reservoir of ore constituents must have been in the Athabasca sandstones and that the basement contributed little. This contention is supported by the observation that below the unconformity, the alteration haloes show a rapid downward narrowing, demonstrating that only a small volume of basement rocks was affected by alteration processes (Figures 8,9,10,11,

13,15). We submit that ore constituents, including U, Ni, Co, Mo, V, Ag, Au, As, Se and possibly minor Cu, Pb, Zn, have been derived from a lateritic palaeoweathering profile and carried into the basin adsorbed to detrital ferric-iron hydroxides and kaolinite. During diagenetic recrystallization of the clay and conversion of limonite into specular hematite, these constituents were released to the pore fluid, thus becoming available to form ore deposits long after deposition of the sediments, during an advanced stage of basin evolution. The basement contributed K, Mg, Fe, now tied up in the alteration haloe, plus hydrocarbons and carbon dioxide and possibly small quantities of base metals. Had basement rocks been extensively leached, base metals liberated from the destruction of ferromagnesians and feldspars might be expected to have been more prominent among the constituents of the unconformity-type deposits. Had feldspars and ferromagnesian minerals been major detrital components of the Athabasca Group rocks, base metals would have been available in quantity from that source. The lack of base metals in the unconformity ores may be a reflection of the mineralogical maturity of the Athabasca basin fill.

The source of sulphide presents a problem of its own. Sulphate may be expected to have been present in the Athabasca sandstones in concentrations well over those of arsenate (Wedepohl, 1978), but it is unlikely to have undergone significant inorganic reduction to sulphide at the prevailing temperatures of 150-200°C (Malinen and Khitarov, 1969; Spirakis, 1981; Omoto and Lasaga, 1982). Bacterial reduction under conditions of deep burial and high-grade diagenesis does not seem likely either. The preponderance of arsenide over sulphide during the first stage of mineralization may reflect a low sulphide availability, arising

from the difficulty to reduce sulphate inorganically at low temperatures.

During the second stage of mineralization the role of the basement as a source of ore constituents may have been somewhat greater. The fact that at the Midwest deposit, basement ore dated at ∿ 1100 Ma (Worden et al., 1984; Badsgaard et al., 1984) represents the second stage of mineralization, may suggest that fluid circulation during this stage reached more deeply below the unconformity than during the first. The sulphide characterizing the second mineralization stage may have been derived from sulphides present in the graphitic pelites, mainly pyrite, some chalcopyrite and sphalerite, as may accompanying base metals.

If a sandstone source is accepted for the ore metals, then the operation of convective flow in the basal Athabasca aquifer becomes a necessary component of our diagenetic-hydrothermal concept, in order to transport dispersed ore constituents to the site of mineralization. As the overall period of mineralization is coeval with the episode of diabase intrusion, we infer that convection within the sandstone may have been driven by increased heat flow associated with diabase magmatic activity (Hoeve and Sibbald, 1978, 1979; Hoeve et al., 1980).

The stability criterion for free, thermal, convection in a horizontal aquifer is expressed by the Rayleigh number (Combarnous and Bories, 1975; van der Kamp, 1984; Wood and Hewitt, 1982, 1984).

$$Ra = \frac{g\alpha(\rho C)_f k \Delta T H}{\mu \lambda^*}$$

in which

g = gravity acceleration

(ρc) = heat capacity of fluid

= thermal expansivity

K = permeability

T = temperature

H = thickness of aquifer

μ = fluid viscosity

λ^* = thermal conductivity of aquifer

The critical value for the Rayleigh number is $Ra_c = 40$. For values of $Ra \leq 40$ the aquifer will remain stagnant, whereas for $Ra > 40$ free convection will be induced, resulting ideally in a series of regularly spaced, hexagonal cells, whose radius is approximately equal to their height.

For the Athabasca basin we can make the following estimates for the parameters in the equation

H - corresponds to the thickness of the Manitou Falls Formation (MFa + b plus MFc + d), assuming the silt- and clay-rich Wolverine Point Formation formed an upper seal to the aquifer; therefore, $H = 1000$ m, depending on location within the basin.

ΔT - temperature difference between top and bottom of the aquifer. For $H = 1000$ m, ΔT corresponds to the value of the geothermal gradient; fluid inclusion evidence suggests a gradient in the order of 35°C/km (Pagel, 1975).

The permeability is assumed to be $k = 50$ md, which is a fairly typical value for buried sandstones (Loucks et al., 1984). For the

remaining parameters, published literature data for concentrated (25-30% NaCl) brines were used (Smith, 1963; Pryor, 1971; Potter and Brown, 1977).

Entering these estimates into the equation and taking values of 30°C/km, 32°C/km and 35°C/km for the geothermal gradient, Rayleigh numbers of Ra≅33, Ra≅40 and Ra≅46, respectively, are produced. Thus, for geologically reasonable conditions, convection in the basal Athabasca aquifer may have been controlled by temporal variations in heat flow. This is an important result, because it provides for a switching mechanism to regulate convectively driven diagenetic-hydrothermal systems over the life of the basin, on a basin-wide scale and in tandem with episodic diabase magmatism, tectonic reactivation and active deepening of the basin. Such a mechanism may account for the delayed start-up of ore-forming systems under conditions of deep burial some 100-150 Ma after deposition of the sediments and for the extended 50-100 Ma duration of each of the main stages of hydrothermal activity.

The initiation of convection and the spatial distribution of cells may be expected to have been strongly influenced by the presence of graphite in basement rocks, because even small admixtures of graphite would raise the conductivity of a rock by orders of magnitude due to the very high thermal conductivity of this mineral (graphite: 400-1000 x 10^{-3} cal/cmsec°C; feldspar: 10 x 10^{-3} cal/cmsec°C; Clark, 1966). As pointed out by Tilsley (1980) in a different context, thermal gradients in graphite-rich rocks may be as low as 1-2°C/km, so that the temperature of a graphitic zone at the unconformity may be essentially equal to the ambient temperature at depth. For a steeply dipping graphite-rich zone extending 1.0-1.5 km below the unconformity, the temperature at the

unconformity may be estimated to have been 30°-50°C higher than of adjacent, non-graphitic, quartzofeldspathic gneisses, resulting in a locally increased T and a corresponding increase in the value of Ra. This means that starting from a condition of stagnancy ($Ra \leq 40$) and in a situation of increasing heat flow, convection in the Athabasca aquifer would tend to be focussed upon graphitic zones at the unconformity. Consequently, we expect that virtually every graphitic zone sub-cropping at the unconformity will have been associated with a couplet of convective cells and that the configuration of cells within the basal Athabasca aquifer will have been strongly affected by the distribution of graphitic rocks in the basin floor. The long linear shapes of the ore bodies of the Key Lake, Collins Bay B, Midwest, McLean Lake and Cigar Lake deposits (up to 2000 m long and 30-50 m wide) suggest that the actual cells may have developed into horizontal longitudinal rolls rather than regular hexagons. Geological evidence of raised temperatures in graphitic rocks is provided by quartz dissolution and residual enrichment of clay within host rock alteration haloes, both immediately below the unconformity and above it in the sandstone (Figure 15). The inferred mechanism of focussed convection is analogous to that of a pan of water coming to a boil. Although convection in the water is driven by the overall heat flow through the bottom of the pan, cells focus on hot spots that actually touch the heating element underneath.

In the absence of plumbing system allowing Athabasca pore fluids access to the underlying basement rocks, convection in itself would seem unlikely to promote significant diagenetic-hydrothermal exchange between sandstone and basement, because the basin floor will have been essentially sealed. Contemporaneous faulting, therefore, constitutes an ad-

ditional critical factor in our model. Faulting, accompanying episodic tectonic reactivation and deepening of the basin may be expected to have been focussed upon graphitic layers in the basement on account of the lubricant properties of graphite. Thus, graphite may have exerted a dual control on the location of diagenetic-hydrothermal systems of sandstone-basement interaction (1) a thermal control by virtue of its conductive properties and (2) a mechanical control by virtue of its lubricant properties. In addition graphite exerted a chemical control on the mineralization process by generating mobile hydrocarbon reductants from its reaction with incoming pore waters.

Attempts to evaluate the chemical conditions of ore formation are hampered by the lack of thermodynamic data for Ni-Co-arsenides. Nonetheless, a rough estimate can be made from the observed mineral assemblages and alteration features. The log fO_2-pH diagram shown in figure 18 was calculated using thermodynamic data of Barnes and Crerar (1976) and Helgeson (1969). At the point of discharge, the composition of the reduced fluids emerging from the basement, must have plotted approximately at point A of the diagram, because host rock alteration is marked by illitization (chloritization), graphite destruction and hydrocarbon formation. Absence of siderite during the first stage of mineralization puts an upper limit to $\Sigma C = 10^{\circ}$ for $a^{Fe^{2+}} = 10^{-5}$. At the zone of secondary hematite both illite and hematite were stable so that the fluid composition here is approximately represented by point B. The field of pyrite stability is assumed to approximate the field of ore precipitation, even though the first stage of mineralization is dominated by arsenides rather than sulphides. In ore-forming systems, mixing of the emerging reduced fluids with oxidized Athabasca pore fluids created a

regular, very high redox gradient and a very low pH gradient indicated by the path A-B-C. It appears that two reaction fronts developed (1) a redox front controlling the precipitation of ore (point B), and (2) a pH front controlling clay alteration (point C). Consequently, ore formation was not directly coupled to clay alteration as had been suggested for the Jabiluka deposit in Australia (Gustafson and Curtis, 1983).

The second phase of mineralization is characterized by extensive bleaching and by precipitation of siderite and base metal sulphides, suggesting a lower activity of ferrous iron in solution and a higher activity of carbon e.g. $a^{Fe^{2+}} = 10^{-7}$ and $\Sigma C > 10^{\circ}$ (figure 17).

The presence of extended wings to the clay alteration and quartz dissolution haloes laterally along the unconformity (Figures 9,10,11,13, 16) may appear somewhat enigmatic as it seems to argue against our contention that the diagenetic-hydrothermal systems of sandstone-basement interaction are driven by convection in the Athabasca aquifer. Although we have not attempted to quantitatively model the Athabasca flow systems, a comparison with hydrothermal circulation at mid-ocean ridges may be instructive. Parmentier and Spooner (1978) have modelled the convection of sea water involved in the formation of massive sulphide deposits at Troodos. The configuration of flow lines and isotherms as computed by Parmentier and Spooner (1972) for a Rayleigh number Ra = 50 in a medium of uniform permeability and for variable, temperature dependant, fluid viscosity (μ) and compressivity (α), is reproduced in Figure 19. Notable aspects are the generation of a narrow thermal plume at the axis of the cell and a distinct bunching of isotherms parallel to and just above the base of the cell. For the Athabasca basin the situation would differ by a lack of fluid exchange across the top of the cell

(relatively impermeable Wolverine Point formation) and by non-uniform heat flow across the base of the cell on account of the presence of graphitic conductors in the basin floor. Closure of flow lines at the top of the cell and a possible enhancement of the axial thermal plume may be expected among the effects of these differences. Assuming a thermal structure analogous to that depicted by Parmentier and Spooner (1978), a schematic cross-section of a convective system such as envisaged for the Athabasca basin is shown in Figure 19. We feel justified in making this assumption because the thermal structure of the Athabasca hydrothermal systems as portrayed by the quartz dissolution contours (Figure 15) is similar to that modelled by Carpentier and Spoover for mid-ocean ridges. Realizing that for a given activity of potassium in solution the stability of illite is strongly dependant on temperature, the formation of the laterally extended wings to the alteration haloe may be due to a thermal shift in illite stability brought about by the bunching of isotherms just above the unconformity. For a given activity of potassium, a temperature increase of 50°C (e.g. T=170° to T=220°C) will shift the kaolinite/illite boundary by 0.25 units towards lower pH values (Figure 17). Hence, in a situation where the potassium activity is buffered by a diagenetic kaolinite/illite assemblage, convective flow in a positive thermal gradient may induce illitization of kaolinite and transfer of potassium from areas of lower to areas of higher temperature. As argued by Wood and Hewett (1984), the converse would be the case for quartz, because this mineral is more soluble at higher than at lower temperatures. The strongest mineralogical effects may be expected in areas of rapid temperature change, such as in the region of bunched isotherms near the unconformity and within the thermal plume. If this

interpretation holds true, the formation of the clay alteration haloe would be due to the dual effects of thermal convection and the discharge of potassium-bearing fluids from the underlying basement.

The overall process that led to host rock alteration and mineralization can now be reconstructed as follows (Figure 19). Following deep burial and under conditions of high-grade diagenesis, free convection was induced in the basal Athabasca aquifer at approximately 1350 Ma by increased heat flow associated with the onset of an episode of tectonic reactivation, diabase magmatic activity and active deepening of the basin. Convective cells were focussed on graphitic layers subcropping at the unconformity, because the temperature of these layers was in the order of 30-50°C higher than the ambient temperature at the floor of the basin. Since the graphitic layers constituted linear thermal anomalies, couplets of longitudinal convective rolls formed alongside the trace of the graphitic rocks, rather than regularly spaced hexagonal cells. Diagenetic-hydrothermal sub-systems of sandstone-basement interaction developed throughout the Athabasca basin, wherever graphitic zones were fractured by contemporaneous faulting, thus allowing sandstone pore waters access to the basement rocks underneath. Assuming equilibrium between pore waters and the hematite-kaolinite-illite mineral assemblages of the basal Athabasca sandstone (Manitou Falls Formation), the composition of the diagenetic fluids may be expected to have plotted on the kaolinite/illite boundary of the diagram shown in Figure 18 (e.g. point C). Where such fluids gained access to the basement via fault zones intersecting the basin floor they will have reacted with the predominantly quartzofeldspathic and pelitic metamorphic rocks. Reaction with graphitic metapelites, leading to graphite destruction and the

formation of solid hydrocarbons, transformed the incoming oxidized fluids into reduced fluids containing hydrocarbons and carbon dioxide. The temperature of the incoming fluids was raised in the order of 30-50°C by contact with graphitic basement rocks, rendering the reduced fluids undersaturated in silica. Illitization of K-feldspar in basement rocks shifted the composition of the fluids away from the kaolinite/illite boundary into the illite field owing to loss of H^+ and gain of K^+ incurred in the process. Similarly, destruction of plagioclase and chloritization of mafic minerals may be expected to have contibuted sodium, calcium, magnesium and ferrous iron to the fluid, plus possibly small quantities of base metals. Where Athabasca pore fluids gained access to dolomitic basement rocks as at Rabbit Lake, they induced widespread carbonate dissolution resulting in collapse brecciation. The prevalence of magnesium in this environment produced chloritic rather than illitic host rock alteration. Upon re-emergence above the unconformity, the modified fluids will have differed significantly from the prevailing Athabasca pore waters in that they were at a higher temperature, less dense, undersaturated in silica, enriched in potassium, magnesium and ferrous iron, strongly reduced, more alkaline and, therefore, no longer in equilibrium with the hematite-kaolinite-illite-quartz assemblage of the sandstone.

At centres of discharge, quartz was dissolved from the rocks within the thermal anomaly created by graphitic rocks at the unconformity, giving rise to a haloe of residual clay enrichment and the formation of a distinct clay cap over the graphitic zone. The configuration of palaeoisotherms is depicted by the contours outlining the disappearance of quartz overgrowths (Figure 15). In ore forming systems, mixing with

prevailing, oxidized, Athabasca pore waters produced a regular redox gradient outwards from the point of discharge (Figure 19). Ferrous iron carried by the reduced fluids remained dissolved until at some distance from the discharge point a specific redox value was exceeded where it was precipitated as hematite, giving rise to the zone of secondary hematitization which marks the position of a distinct and stationary redox front. Below this front, diagenetic, specular hematite was destroyed through chemical reduction of ferric iron; above the redox boundary, purple, diagenetic hematite was preserved. Ore constituents carried by the oxidized pore waters were precipitated below the redox front, probably according to a zonal pattern in response to the prevailing redox gradient. At the McClean deposit, a regularly increasing U/Ni ratio outwards from the core of individual ore pods has been interpreted as to reflect such a regular redox gradient (Wallis et al., 1983). A similar zonation has been recognized at the Collins Bay B, Midwest and Cigar Lake deposits (Jones, 1981; Salaff, 1984; Ruzicka, 1984), as has a crude mineral zonation pattern in which tri- and diarsenides in the core give way externally to sulpharsenides and minor Ni-sulphides.

Concomitently with redox reactions leading to the formation of the high-grade ore body, pH controlled reactions gave rise to the conversion of diagenetic kaolinite into illite and chlorite, forming an illitization/ chloritization haloe around the point of discharge. In the process the composition of the fluids shifted back towards the kaolinite/illite boundary until in equilibrium with the interstitial mineral assemblage of the unaltered sandstone at some distance from the discharge point. Because the conversion of kaolinite into illite is unaffected by the redox gradient, the illitization front need not coincide

with the redox front, and is indeed seen to be located far beyond the zone of secondary hematitization. The laterally extended wings to the alteration haloe just above the unconformity formed in response to the thermal regime created by convection in the Athabasca aquifer.

In this reconstruction the dissolution of quartz within the alteration haloe is strictly a consequence of the thermal regime imposed by the graphitic rocks and the convective flow in the aquifer, and is independent of both clay alteration and mineralization. Furthermore, the mineralogical character of the alteration haloe i.e. illitization or chloritization, depends entirely on the composition of the basement rocks that are being leached: predominantly illitization, where the basement consists of quartzo-feldspathic rocks as at Midwest, McClean, Key Lake; predominantly chloritization, where the basement consists of magnesium-rich marbles and calc-silicates as at Rabbit Lake. The establishment of a stationary redox front during the first stage of mineralization is of critical importance to the formation of the ore body. Without it, ore precipitation, if taking place at all, will not have been focussed and may be expected to have been highly dispersed, resulting in the formation of clay mineral haloes lacking significant mineralization. In actual case, numerous barren clay mineral haloes which are known in the Athabasca basin, may reflect exactly such a situation. The alteration characteristics of such haloes, as far as we know, are similar to those of ore-bearing haloes, including quartz dissolution, development of a residual clay zone just above the unconformity and in the uppermost basement, as well as superimposed bleaching representing the second stage of hydrothermal activity. Despite all this, mineralization is dispersed and insignificant to non-existent. In one example from the

northwestern portion of the basin (confidentiality requirements preclude disclosure of location and presentation of a cross-section), extensive bleaching forms a well-defined haloe of several hundred metres in diameter which is approximately congruent with a clay mineral haloe. The only evidence of mineralization is in the form of small, patchy, accumulations of ore minerals that are scattered along the interior, reduced, margin of the bleaching/clay alteration front against purple, hematitic sandstone. A stationary redox front, however, does not seem to have developed, as there is no evidence of a zone of secondary hematitization.

A similar situation may have prevailed during the second stage of mineralization at 1100-1050 Ma. Although large-scale bleaching attests to the generation of reduced fluids and basin-wide reactivation of diagenetic-hydrothermal systems, fluid discharge took place under a different hydrodynamic regime and no stationary redox front developed.

Within the context of our model of sandstone-basement interaction, a potential for mineralization not only exists at the discharge end of diagenetic-hydrothermal flow systems such as discussed above, but also at the recharge end where oxidized Athabasca pore fluids enter the basement, provided the influx is sufficiently focussed. Examples of the latter situation may possibly be presented by Rabbit Lake, some of the Cluff Lake deposits and the little-known Eagle Point deposit near Collins Bay. At Eagle Point, fractures in slightly graphitic basement rocks are flanked by, successively, an internal envelope of hematitic clay alteration next to the fracture and an external envelope of pale green alteration (chlorite plus illite) farther away. Pitchblende mineralization is located at the redox interface between hematitic and

non-hematitic altered rock. For the Cluff lake N-zone and Claude deposits, Harper (1980) reports a spectacular mineral and element zonation pattern in which a pitchblende zone, again located at the redox interface, gives way successively to a distinct jordisite front (expressed as molybdenum bloom) and a zone containing pyrite and solid hydrocarbon buttons. The patterns of hematitization and ore zonation resemble those observed in sandstone-type deposits (Harshman, 1974) and attest to the passage of oxidizing, metalliferous fluids through the basement rocks. At Rabbit Lake the extensive dissolution of carbonate, the presence of collapse breccias and the chloritization and hematitization point to a influx of corrosive, oxidized ore fluids into the basement.

As a group these basement-located deposits lack the characteristic association with abundant Ni-Co-arsenides of the Midwest, McLean, Key Lake and Collins Bay deposits, which formed at the discharge end of diagenetic-hydrothermal flow systems of cover-basement interaction. Possibly, these differences may reflect less reducing conditions in the formation of the basement-hosted, influx-related deposits.

In the scenario outlined above, the establishment of diagenetic-hydrothermal systems of sandstone-basement interaction and the formation of ore are local manifestations of a much broader process of mass transfer, chemical differentiation and diagenesis within large-scale convection systems. For example, quartz leached from the warmer core regions of the convective flow systems will have been carried upwards and reprecipitated as authigenic overgrowths around detrital quartz, thus contributing to sandstone diagenesis. Potassium, precipitated as illite in the lateral wings to the alteration haloe just above the unconformity, has probably been derived from cooler, upper regions of the convective

cell. Ore constituents inferred to have been released to the Athabasca pore fluids during diagenetic recrystallization of limonite into hematite, were transported to and trapped by diagenetic-hydrothermal systems and precipitated as ore. Thus, distinctions between diagenesis, host rock alteration and mineralization become meaningless, as they are merely different aspects of one and the same underlying process of convection-mediated mass transfer in the basal Athabasca aquifer. This notion is supported by recent studies (Tardy and Cassan, 1981; Wood and Hewett, 1982, 1984; Galloway, 1984; Blanchard and Sharp, 1985) which suggest that free, heatflow-controlled, convection may be a common feature in hydrocarbon reservoir rocks, and that convective flow is an essential factor in burial diagenesis. Even with typical fluid velocities as low as 1 m/year, free convection is considered capable over geologic time to effect significant mass transfer resulting in, for example, quartz and carbonate cementation of permeable sandstones and in development of secondary porosity in spatially separated regions of large-scale convective cells. This leads Wood and Hewett (1982, pp. 1707) to suggest that many sandstone reservoirs can be considered "as low-grade hydrothermal systems" and that "diagenesis can be regarded as a low-heatflow-analogue of hydrothermal alteration, occupying the low-fluid-flow regime of the spectrum passing into the hydrothermal regime as heat flow and fluid velocities increase".

This quotation applies to the situation in the Athabasca basin, which we believe to provide an example of a basin in which diagenesis and hydrothermal mineralization are coupled by free convection. The term "diagenetic-hydrothermal", coined earlier (Hoeve and Sibbald, 1976,

1978) to describe our metallogenetic model, aptly expresses these relationships.

SIGNIFICANCE TO URANIUM EXPLORATION

Unconformity-type uranium ore bodies in the Athabasca basin tend to be of high grade and very small dimensions in the order of 10-100 m wide and up to 1000-2000 m long, which makes them difficult targets for exploration, particularly when covered by a considerable thickness of Athabasca sandstones and conglomerates. The development of criteria that allow enlargement of potential drill targets is therefore of obvious importance.

In this contribution we have shown that host rock alteration features such as clay alteration, bleaching and quartz dissolution are diagnostic of the mineralization process and also constitute haloes that extend well beyond the limits of the actual ore bodies. Host rock alteration is present along the full strike length of ore-bearing fault zones, albeit of varying intensity and extent, and major ore accumulations are associated with distinct alteration pipes that rise several hundred metres above the unconformity up to the level of the present bed rock surface.

From a practical point of view these are significant characteristics, because they allow host rock alteration to be applied as drill hole evaluation and ore guide criteria in exploration programs. We estimate that host rock alteration criteria may enlarge potential drill targets by a factor of 10-20. Information on the configuration of bleaching haloes can be obtained from systematic core logging; diagenetic clay mineral alteration is readily identified by means of automated quantitative XRD techniques, whereas quartz dissolution and residual clay enrichment can be quantified by simple gravimetric analysis of total clay content. Additonal criteria for the recognition of diagnos-

tic clay mineral alteration are provided by the illite (002)/(001) peak intensity ratio (Table 9) and by illite polytype. Alteration illites have a higher I(002)/I(001) ratio than diagenetic illites in the adjoining unaltered sandstones and, in addition, are characterized by the 3T polytype as opposed to 2M and 1M polytypes in unaltered sandstones.

The fact that host rock alteration extends for several kilometres along strike of ore-bearing faults suggests that the mineralization potential of a given conductive zone can be quickly evaluated by a limited number of widely spaced drill hole fences, with intervals in the order of 2.0 km. In areas where the thickness of Athabasca cover is less than approximately 400-500 m, shallow bed rock samples may be utilized to outline the surface expression of alteration pipes associated with major centres of hydrothermal activity i.e. to identify specific targets for diamond drilling (Figure 12).

During evaluation drilling of known occurrences host rock alteration may be used as a guide to ore. In this application a third hydrothermal centre was indicated for the Midwest deposit, located between the two known ore bodies, which as yet is untested for ore (Hoeve, 1982, 1984).

At Key Lake the situation is more complex than at the other deposits of the basin owing to superimposed kaolinitization associated with the late 300-250 Ma remobilization of the ore. Since the kaolinitization is not diagnostic of the primary ore-forming process, it cannot be used directly to identify centres of hydrothermal activity. Nonetheless it is useful in outlining fault zones in the sandstone, particularly when analyzing top bedrock samples. Recognition of quartz dissolution

haloes on the basis of increased total clay content could still identify targets for diamond drilling.

Another complicating factor is that in the "Greater Key Lake Region" individual alteration haloes spreading from several hydrothermal centres have apparently merged so as to produce a semi-regional alteration blanket in the basal sandstone. In this situation potential targets become too large, so that other alteration features such as the illite peak intensity ratio I(002)/I(001), in conjunction with quartz dissolution and residual clay enrichment may provide better criteria to locate centres of hydrothermal activity than proportions of clay mineral species alone.

The exploration experience in the Athabasca basin is that many, if not all, graphitic layers sub-cropping at the unconformity are associated with at least weak alteration in the immediately overlying sandstones, and further, that several large alteration haloes are essentially barren. If we are correct in our assessment that the establishment of a stationary redox front is a "conditio sine qua non" for the formation of high-grade ore this circumstance places a caveat on the application of alteration criteria to exploration. Although host rock alteration features have proven to be very helpful in identifying centres of diagenetic-hydrothermal activity, they cannot as yet predict the presence of ore.

COMPARISON WITH AUSTRALIAN DEPOSITS

The unconformity-type deposits of the Athabasca basin have much in common with the uranium deposits of the Alligator River Uranium Field in the Pine Creek Geosyncline of the Australian Northern Territories. Deposits in the two districts share characteristics of geological setting and age, aspects of host rock alteration, a close spatial association with a major Middle-Proterozoic unconformity and, most significantly, a general association with graphite-bearing basement rocks (Nash et al., 1981). Radiometric studies of the Australian deposits suggest two main episodes of post-Kombolgie mineralization and remobilization, at 1600 Ma and 900 Ma, respectively (Hills and Richards, 1976; Ferguson et al., 1980; Page et al., 1980; Ewers et al., 1983). Destruction of graphite within bleached haloes around pitchblende veins is reported for some deposits (Binns et al., 1980), as is the local presence of solid hydrocarbon matter which is occasionally trapped within fluid inclusions enclosed in ore stage quartz (Ypma and Fuzikawa, 1980; Ypma presented a slide showing such an inclusion at the Pine Creek Geosyncline Uranium Symposium, Sydney, 1979). The similarities between the two districts are particularly striking for carbonate-hosted deposits such as Jabiluka and Ranger in Australia and Rabbit Lake in Saskatchewan, which are characterized by extensive carbonate dissolution of dolomitic and magnesitic marbles of the metamorphic basement, collapse brecciation, chloritization and chert replacement of carbonate (Eupene et al., 1975; Hegge and Rowntree, 1978; Hoeve and Sibbald, 1978; Gustafson and Curtis, 1983).

At Jabiluka, Gustafson and Curtis (1983) identified widespread chloritization accompanied by extensive quartz corrosion, carbonate

veining and younger quartz veining in the Kombolgie sandstone overlying the deposit, forming a haloe of at least 300 m in diameter. The chloritization of the sandstone appears to be identical to the chloritization of mineralized basement rocks, as attested to by enrichment of a similar trace element suite and by identical chlorite compositions. Ore is, however, limited to the basement where it is primarily located in zones of intensive chloritization and collapse brecciation. The dissolution and collapse of the carbonates must have taken place after lithification of the Kombolgie Formation, as rotated, angular blocks of coherent sandstone are incorporated within collapse breccias below the unconformity. Extensive chert replacement of carbonate, accompanied by a significant volume loss, reported for the Jubiluka and Ranger deposits (Eupene et al., 1975; Hegge and Rowntree, 1978), is probably equivalent to secondary hydrothermal quartzite at Rabbit Lake, whose formation was interpreted as being due to reprecipitation of silica liberated in the chloritization of pelitic and quartzo-feldspathic rocks intercalated with the carbonates (Hoeve and Sibbald, 1978).

The overall characteristics of the Australian and Saskatchewan deposits are so similar that a common origin may reasonably be assumed, so that metallogenetic models developed for one district should also be applicable to the other. Early genetic models for both the Australian and Saskatchewan deposits emphasized the spatial proximity of the palaeo-weathering profile and called upon surficial weathering and erosion-related processes as principal agents of ore formation (Knipping, 1974; Langford, 1974; McMillan, 1978; Ferguson et al., 1980; Tilsley, 1980). Later a polygenetic origin was advocated for deposits in both districts, involving syngenetic enrichment in lower Proterozoic

sediments, followed by progressive upgrading during regional metamorphism, pre-Athabasca/Kombolgie erosional enrichment, post-Athabasca/Kombolgie remobilization and final concentration (Dahlkamp, 1978; Kirchner et al., 1980; Pagel et al., 1980; Dahlkamp and Adams, 1981; Tremblay, 1982; Ewers et al., 1983).

The early consensus between Australian and Saskatchewan geologists collapsed when it was realized that, at least for the Saskatchewan deposits, the pre-diagenetic part in the reconstructed sequence of events leading to the formation of ore, was entirely conjectural (Hoeve and Sibbald, 1976; 1978; Hoeve et al., 1980). Since then diagenesis of the overlying red-bed sandstones has been recognized to also constitute the principle factor in the formation of the Australian unconformity-type deposits. On the basis of fluid inclusion studies Durak et al., (1983) pointed to compositional similarities between fluids involved in Kombolgie diagenesis and those involved in host rock alteration. Gustafson and Curtis (1983) presented a genetic model for the Jabiluka deposit that is essentially similar to the diagenetic-hydrothermal model presented by Hoeve and Sibbald (1976; 1978) for the Athabasca deposits. Although Gustafson and Curtis (1983) do not specify interaction between oxidized and reduced fluids, their model is characterized by penetration of Kombolgie formation waters below the unconformity; by reaction of these waters with basement rocks leading to carbonate dissolution, collapse brecciation and chloritization; by precipitation of uranium from these fluids due to reduction by graphite and by re-emergence of the fluids above the unconformity to form the chlorite haloe in the overlying sandstone. They state (pp. 26) that "The Kombolgie sand-

stones were an essential part of the formation of the high-grade ore bodies...", and further suggest that, as in the Athabasca basin, the hydrothermal system was driven by a regional heating event associated with the intrusion of phonolite and diabase dikes.

The main difference between the Australian and Saskatchewan deposits is that in the Alligator Rivers Uranium Field mineralization appears to be restricted to basement rocks, whereas in the Athabasca basin ore is primarily found at and above the unconformity. This may be, however, not so much an inherent difference between districts as a reflection of the effects of host rock lithology on the development of the hydrothermal plumbing system, considering that the closely similar deposits of Rabbit Lake, Midwest, Jabiluka and Ranger are hosted by carbonates, whereas those of Key Lake, Midwest, Cluff Lake and McClean Lake are underlain by quartzo-feldspathic and pelitic rocks. If Kombolgie diagenetic fluids may be assumed to have had a composition and pH comparable to those of the basal Athabasca aquifer (Figure 17), which were buffered by the assemblage kaolinite/illite/hematite, then these pore fluids must have been under-saturated with respect to Ca-Mg-carbonates. Where such fluids penetrated the basement and encountered dolomitic or magnesitic marbles they will have dissolved the walls of the initial fractures, thereby increasing permeability and promoting increased flow. The result would be a highly permeable, self-propagating plumbing system, which would follow stratigraphy and which eventually could reach very deep below the unconformity. This may be the reason for the large dimensions of basement alteration haloes in carbonate-hosted deposits, as compared to the rapidly downward-narrowing, funnel-shaped haloes in deposits hosted by pelitic and quartzo-feldspathic rocks (Figure 15).

It still leaves the question, however, why in deposits such as Jabiluka, which is overlain by up to 300 m of sandstone, virtually no ore is found above the unconformity. Yet a potential for mineralization by a mechanism of hydrodynamic and chemical interaction of reducing and oxidizing fluids must have clearly existed at Jabiluka, because the presence of the alteration haloes in the sandstone above the deposit implies discharge of Mg-rich, reduced, fluids above the unconformity. We suggest that the absence of sandstone and unconformity ore at Jabiluka may possibly reflect a hydrodynamic situation in which a stationary redox front failed to be established above the unconformity, analogous to that inferred for the formation of barren alteration haloes in the Athabasca basin.

OTHER BASINS AND OTHER METALS

In a broader perspective, the results of the present study may be of significance to mineralization in other sedimentary basins and to metals other than uranium. As there is no evidence to suggest that the Athabasca and Kombolgie basins are unique in any respect, it is reasonable to suppose that similar processes have been operating in other sedimentary basins containing sections of permeable sandstones. If free convection is indeed as common in deeply buried sandstones as proposed in recent publications (Tardy and Cassan, 1981; Wood and Hewett, 1982; 1984; Galloway, 1984), this would suggest that in the course of basin evolution the interplay of burial diagenesis, convective flow and cover-basement interaction may be of much greater importance in the formation of a variety of unconformity-associated and sediment-hosted ore deposits than is presently realized. This would be especially true for metal deposits associated with thick sections of red beds, since the association of uranium, copper, lead and vanadium mineralization with red beds has long been recognized (Chandler, 1980) and a genetic link with red bed diagenesis suspected (Holmes et al., 1983). In stratigraphic settings such as that of the Athabasca basin, a potential for mineralization would exist not only at the unconformity, but also elsewhere in the sedimentary sequence wherever suitable reducing traps happen to be available, for example, in the form of intercalated reduzate shales. The composition of the resulting ores would to a large degree depend on the chemical and mineralogical composition of the red bed source rocks and on their diagenetic history. This may be illustrated by the following examples taken from the literature.

Other unconformity-type uranium deposits

In the Russian literature two uranium deposits are reported from undisclosed locations in the Soviet Union, which share the characteristic features of the Canadian and Australian unconformity-type deposits. Volshin and Korolev (1982) describe a deposit in which uranium mineralization forms narrow, linear ore bodies which straddle the unconformity between a folded, Cambro-Ordovician basement, affected by Caledonian orogenesis, and overlying Middle-to Upper-Devonian red beds. Mineralization is fault-controlled and occurs within 100 m from the unconformity. Host rock alteration is represented by silicification, quartz-sericite-pyrite alteration, secondary hematitization and by chemical reduction of red beds. Pitchblende is associated and intergrown with hematite and iron-chlorite. Barite is present in pre-ore veinlets, whereas ankerite and sulphides (pyrite, galena, arsenopyrite) post-date the uranium mineralization. Temperature of formation as inferred from mineral stabilities and fluid inclusions, is estimated at 150°C. Volchin and Korolev ascribe the origin of the deposit to interaction of oxidized pore waters of the Devonian red beds with ascending waters, carrying gaseous reductants, emerging from faults in the underlying basement.

For another deposit Oparysheva et al. (1974) report that uranium mineralization can be traced from the sedimentary cover into the underlying crystalline basement. Neither the age of mineralization nor that of the host rock are disclosed. The mineralization is fault-controlled and straddles the unconformity between granite basement rocks and overlying clastic sediments, some of which are red beds. The unconformity

is underlain by a limonitic palaeoweathering profile of lateritic character. Host rock alteration, superimposed upon the palaeoweathering profile is characterized by chemical reduction of limonite and hematite, by carbonatization, sulphidization, bituminization, secondary hematitization, clay alteration (kaolinite and illite), silicification and by late superimposed limonitization. Uranium ore forms flat-lying lenses in the cover sediments and fault-controlled veins and impregnations in the granitic basement. Pitchblende is the main uranium mineral and is accompanied by pyrrhotite, marcasite, pyrite, chalcopyrite, sphalerite, galena and jordisite. Based upon fluid inclusion studies, the temperature of formation is estimated at 220-230°C. The origin of the mineralization is attributed to ascending solutions which, emerging from basement faults that intersect the unconformity, were discharged into the overlying sediments.

Unconformity-type lead deposits

Possible examples of unconformity-type lead deposits associated with red beds are provided by several Proterozoic vein deposits in Canada and by the much younger sandstone lead deposits of the northern Eifel, Germany. The Canadian examples include Pb-Zn-Ba veins of the Dorian area, Thunder Bay District, Ontario (Franklin and Mitchell, 1977); Pb-Zn-Cu veins near Artillery lake in the Great Slave Lake region of the Northwest Territories (Gandhi, 1984) and Pb-Zn veins of the Mistassini-Otish basin in Quebec (Chown and Cary, 1973). The geological setting of these deposits is illustrated by the Pb-Zn-Ba veins of the Dorian area, which are spatially associated with the unconformity between an Archean and lower-Proterozoic basement and overlying, Middle-

Proterozoic, unmetamorphosed red beds of the Sibley Group. Temperatures of formation are estimated at approximately 100°C and lead isotopic evidence suggests metal contributions by both basement and arkosic Sibley sandstones. The presence of veins below and above the unconformity is suggestive of fluid interaction between red bed cover and metamorphic basement.

In the northern Eifel, Germany, lead mineralization is represented by the well-known deposits of Mechernich and Maubach and by several smaller occurrences in their vicinity. The mineralization forms dispersed impregnations in Middle-Triassic Buntsandstein red beds and fracture-filling veins in rocks of the underlying metamorphosed Variscan basement. Walther (1984) reports a continuous transition from vein ore in Middle- and Lower-Devonian basement rocks to impregnation ore in the cover rocks. In one deposit, the Kaller Stollen mine, both ore types were exploited. Both ore types contain the same paragenesis, starting with dolomite, siderite, bravoite, quartz and chlorite, followed by sphalerite, galena, chalcopyrite, tetrahedrite, boulangerite, linneaite and millerite (Schachner-Korn, 1960). The temperature of formation is unknown, although the presence of dickite at Maubach (Echle, pers. comm., 1985) would suggest elevated temperatures in the order of 100-200°C. The origin of the deposit has alternately been interpreted as syngenetic and epigenetic-hydrothermal. Bjørlykke and Sangster (1981) classified Maubach and Mechernich as sandstone lead deposits and supported a syngenetic to early-diagenetic origin. Walther (1982; 1983; 1984) considered the deposits as a manifestation of the Saxonian metallogenic epoch, which is assigned a late Cretaceous or Tertiary age on the basis of stratigraphic and structural information.

He advocated an epigenetic origin by ascending hydrothermal solutions, related to tectonomagmatic activity accompanying the opening of the North Atlantic.

If we accept Walther's (1982; 1983; 1984) evidence for a late Cretaceous to Tertiary age for the mineralization, it seems to us that a case can be made for an origin by cover-basement interaction long after deposition of the Triassic host sediments. Lead isotopic evidence indicates metal contributions by both the Variscan basement and the overlying Mesozoic sediments (Large et al., 1983), whereas the continuation of vein ore below the unconformity into impregnation ore above suggests fluid interaction between cover and metamorphic basement.

Unconformity-type Co-As deposit

A different example of a possible unconformity-type deposit is represented by the Co, As (Ni, Fe, Mo, Bi, Cu, Pb, Zn, Ag, Au) deposit of Bou Azzer in Marocco (Leblanc and Billaud, 1982). This deposit consists of a series of ore pods, veins and lodes which are structurally controlled and spatially associated with the unconformity between a lower-Proterozoic basement and overlying Cambrian cover rocks comprising ignimbrites, volcaniclastics and sediments (Figures 1-5 in Leblanc and Billaud, 1982). The ore accumulations straddle the unconformity and are clearly epigenetic; they display mineralogical and chemical zonation and are associated with bleached haloes and zones of potassic host rock alteration. Yet, no magmatic- or metamorphic-hydrothermal activity is recognized in the region. The formation of the deposit is considered enigmatic. Some authors (Routhier et al., 1970; Leblanc and Billaud, 1982) advocate a polystage origin not unlike that initially proposed for unconformity-type deposits (Dahlkamp, 1978; Kirchner et al., 1980;

Dahlkamp and Adams, 1981; Pagel et al., 1980; Tremblay, 1982; Ewers et al., 1983), involving erosion-related pre-concentration followed by a near-endless succession of concentration and upgrading events, all cooperating to ultimately form the present high-grade deposits. Leblanc and Billaud (1982) consider the serpentinite as the source of Co, but have problems in identifying source rocks for As and the other constituents. However, as pointed out by Garcia (1979), the potassic nature of host rock alteration zones flanking mineralized veins implies cover rock involvement, as does the suite of associated elements.

It seems to us that the Bou Azzer deposit may represent another case of diagenetic-hydrothermal interaction between cover and basement in which both basement and cover have chemically contributed to the formation of the deposit. The deposit may illustrate that such interaction is not restricted to red bed cover rocks.

Stratiform copper deposits

Examples of stratiform copper mineralization in a stratigraphic setting very reminiscent of that of the Athabasca basin are provided by the Permian Kupferschiefer copper deposits of Europe and the giant, Proterozoic stratiform copper deposits of eastern Africa, South-Australia and North America (Lur'ye and Gablina, 1973; Rentzsch, 1974; Sawkins, 1976; Gustafson and Williams, 1981; Annels, 1984; Annels and Simmonds, 1984; Brown, 1984). The copper mineralization is characteristically associated with thick sequences of fluviatile, coarse-to medium-grained, oxidized, arkosic sandstones and conglomerates, which were deposited in intracratonic basins, generally in a rift tectonic setting. These are overlain by reduzate marine shales carrying pyrite and organics. The mineralization is spatially associated with basement highs and with the

redox boundary presented by the red bed/shale interface. Fleischer (1984) and Annels and Simmonds (1984) give examples of ore bodies that form semi-annular rings around basement highs capped by dolomitic bioherms against which the ore shales wedge out. In detail the redox front and the mineralization transgress sedimentary boundaries and discordant zonation patterns of metals and minerals parallel to the redox front have been recognized (Lur'ye and Gablina, 1973; Brown, 1974; Rentzsch, 1974; Annels and Simmonds, 1984). The suite of elements associated with the copper includes Zn, Cd, Pb, Co, Ni, Ag, Au, Bi, As, Te, Se, V, U, Mo, Re, although the full suite is rarely present in any single deposit. Contrary to earlier studies which advocate a syngenetic, sedimentary origin for the mineralization (Garlick, 1961; Garlick and Fleischer, 1972; Fleischer, 1972; 1984) recent interpretations favour a post-depositional precipitation of ore metals during diagenesis, although the timing remains highly uncertain between early- and late-diagenetic (Brown, 1971; 1984; Bartholomé et al., 1973; Rentzsch, 1974; Gustafson and Williams, 1984; Knutson et al., 1984). For the White Pine deposit in Michigan, Chaudhuri and Faure (1967) published Rb/Sr and lead isotopic ages of 1075±50 Ma and 720 Ma for, respectively, the Nonesuch Shale host and the chalcocite mineralization, suggesting that mineralization may have taken place during an advanced stage of basin evolution. On the other hand, Ruiz et al. (1984) and Kelly and Nishioka (1985) report a 1047±35 Ma age for the mineralization, based upon a novel Rb/Sr dating method of vein calcite. The source of ore consitituents and the mineralizing fluids continues to be debated. Several authors have pointed to the underlying red beds as a potential source, since these tend to be immature and rich in copper carriers such as mafic minerals and mafic

lithic fragments (Brown, 1971; 1984; White, 1971; Rentzsch, 1974). In a review of sediment-hosted copper, lead and zinc deposits, Gustafson and Williams (1981) argued that the mineralizing fluids were evolved basinal brines, which had leached ore metals from the sedimentary pile in the course of burial and diagenesis. Others sought to relate the polymetallic nature of the associated element suite to the inferred presence of differentiated mafic magmatic rocks deep within the rift structures (Sawkins, 1976; Annels and Simmonds, 1984).

For the copper mineralization of the Keweenaw peninsula, Michigan, Livnat et al. (1983) and Nishioka et al. (1984) indicate a common burial metamorphic origin for (i) chalcocite mineralization in the Nonesuch Shale (White Pine deposit); (ii) traces of native copper, cuprite and copper arsenides in the underlying red-bed Copper Harbour Conglomerate; and (iii) lode and vein mineralization of native copper ± copper arsenides in Keweenawan Portage Lake Lava Series underlying the Copper Harbour Conglomerate. They envision that burial metamorphic fluids of oxidizing character were expelled from the sedimentary pile during deep burial in the Keweenawan rift, and migrated upwards to penetrate the Portage Lake Lava Series, the Copper Harbour Conglomerate and the Nonesuch Shale. Mineral precipitation took place only where suitably reducing traps were encountered by the fluids, as in the lavas and the organic rich Nonesuch shales.

A novel, volcanic, pene-exhalative alternative was proposed by Brown (1984) in which metal-bearing hydrothermal fluids, generated at depth within rift zones, migrated upwards along boundary faults and were injected into the red bed aquifers which thus became charged with ore constituents. Brown suggested that yet-to-be-discovered hydrothermal

feeders and felsic domes might be present at depth near the base of the red bed aquifers.

It seems to us that there is little need to invoke a magmatic or deep-seated hydrothermal origin for the metals, considering that the polymetallic suite is strikingly similar to the one that characterizes unconformity- and sandstone-type uranium deposits for which a magmatic source can be ruled out. We would rather argue that the polymetallic suite is characteristic of the red bed environment, as illustrated by its presence in cupriferous and/or uraniferous reduction spheres in red bed sandstones referred to earlier in this contribution (Harrison, 1975; van de Poll and Sutherland, 1976; Durrance, 1978; 1980; Dyck and McCorkell, 1981), by results of experimental leaching of first-cycle red beds (Zielinski et al., 1983) and by the well-known association of continental red beds with mineral deposits of U, Cu, Pb, and V (±Ni, Co, As, Ag) (Chandler, 1981; Holmes et al., 1983).

The main differences in geological setting between unconformity-type uranium deposits and giant stratiform copper deposits appear to be (1) the presence of a reducing trap at the top of a red bed aquifer rather than at the underlying unconformity, although for the Zambian Copper Belt Mendelsohn (1961) and Pienaar (1961) report that the basement is locally mineralized in areas where ore horizons wedge out against basement topographic highs; (2) the composition of the basin fill. The Athabasca basin fill consists of mineralogically mature, primary red bed sandstones and conglomerates that were derived from a kaolinitic, lateritic weathering profile and deposited under humid, warm climatic conditions, as indicated by the presence of diaspore at several stratigraphic levels. By contrast, the basin fill associated with

stratiform copper deposits consists of immature, arkosic red beds, rich in mafic constituents, which were deposited in a rift tectonic environment under arid, warm, climatic conditions.

Stratiform uranium deposits

For the Athabasca basin, we could speculate that, had the marine Wolverine Point Formation overlying the basal fluviatile sandstones been chemically reducing and rich in organics, stratiform uranium deposits of a diagenetic-hydrothermal origin might have formed, analogous to those in the Proterozoic Franceville basin in Gabon and the Permian Lodève basin in France. In these two basins, stratiform accumulations of uranium are located at the base of organic-rich pelites overlying thick sections of red bed sandstones. The occurrences have been interpreted as typical examples of syngenetic to early-diageneitic mineralization, which had undergone subsequent upgrading to form structurally controlled economic deposits (Chauvet, 1975; Pfiffelman, 1975; Gauthier-Lafaye, 1975; Gauthier-Lafaye et al., 1980; Loversann, 1978; Dahlkamp and Adams, 1981; Ziegler and Dardel, 1984). Recent studies of fluid inclusions, clay mineral diagenesis and readiometric dating, however, failed to uncover any evidence of syngenetic precursors, drawing the assumed sedimentary origin of the mineralization into question. Rather, the studies reveal that in both basins mineralization took place under conditions of high-grade diagenesis (T ~ 200°) during advanced stages of basin evolution (Openshaw et al., 1978; Lancelot et al., 1984).

In the Lodève basin, a U-Pb study of stratiform uranium mineralization identified two stages of epigenetic mineralization at 170 Ma and 110 Ma respectively (Lancelot et al., 1984). The first stage at 170 Ma took place approximately 100 Ma after deposition of the Permian host

sediments, and was coeval with anchizone diagenetic crystallization of illite during a phase of tensional tectonics and associated mafic magmatic activity. The second stage at 110 Ma, although less well-defined, is thought to represent an event of basinal uplift.

For the Franceville basin, Openshaw et al. (1978) conclude on the basis of fluid inclusion evidence that uranium mineralization is tied to high-grade diagenesis, at a temperature of 240°C and a pressure of 1 kb. They interpret the mineralization as a result of interaction of an oxidizing aquifer with overlying organic-rich pelites.

Unifying features

Accepting a red bed source for the ore metals, we suspect that mineralogically mature red beds such as those of the Athabasca basin tend to generate U-Ni-Co-As deposits and that immature red beds may produce base metal deposits. The question whether, in the latter case, copper or lead deposits are formed may depend not only on the composition of the basin fill, but also on the diagenetic history of the immature red beds. The absence of barium from stratiform copper deposits may indicate preferential leaching of mafics, whereas the characteristic association of barium with lead in sandstone lead deposits may point to leaching of feldspars.

We have previously interpreted the Athabasca unconformity-type deposits as a special class of sandstone-type uranium deposits, the latter representing transient stages in a process of red bed diagenesis and basin evolution that has gone to completion in the Athabasca basin (Hoeve et al., 1980). We now propose to extend this concept to stratiform copper deposits, stratiform uranium deposits and sandstone lead deposits. We speculate that the giant stratiform copper deposits may

relate to red-bed type copper deposits in a similar fashion as unconformity-type to sandstone-type uranium deposits. Analogous relationships may exist between syngenetic to early-diagenetic lead deposits such as Largentière and Yava (Samama, 1968; Bjørlykke and Sangster, 1981; Vaillancourt and Sangster, 1984) and unconformity-type lead deposits such as those discussed above.

The diagenetic formation of red beds by means of intrastratal oxidation and alteration starts shortly after deposition (Walker, 1967) and may continue for tens or even hundreds of millions of years, depending on basinal history. Consequently, mineralization resulting from reprecipitation of ore constituents released by the red bed process may also start relatively soon after deposition under near-surface conditons and continue intermittently for a long time in the course of burial diagenesis. In the near-surface environment ore constituents are transported by gravity-driven groundwater flow (Galloway, 1984; Garven and Freeze, 1984) and may precipitate wherever a suitable reducing trap is encountered. Particularly in Phanaerozoic continental sequences which generally contain abundant plant remains that provide numerous reducing traps, the resulting mineral deposits may tend to be small, scattered and of low grade, as is the case for most sandstone-type uranium, red bed copper and sandstone lead deposits (Woodward et al., 1974; Rose, 1976; Chandler, 1980; Ripley et al., 1980; Bjørlykke and Sangster, 1981; Holmes, 1983).

In deeper portions of active sedimentary basins and during advanced stages of basin development, fluid flow is probably driven primarily by compaction and thermal convection (Tardy and Cassan, 1981; Galloway, 1984; Wood and Hewett, 1984). Compaction, by itself, may be expected to

generate single-pass flow systems in which metalliferous fluids pass through the zone of mineralization, precipitate ore minerals and escape from further participation in the formation of the deposit. Although compactional waters are undoubtedly capable of transporting ore constituents, the enormous quantities of fluids required to produce significant ore necessitate very strong focussing of fluid flow (White, 1971; Cathles and Smith, 1981; Sawkins, 1984). Calculations for the White Pine deposit in Michigan show that compactional waters of the red bed Copper Harbour Conglomerate (2 km thick) would have been just barely adequate to generate the deposit, even assuming 50 ppm dissolved Cu (White, 1971). By comparison, saline formation waters (~25% NaCl equ.) from a modern analogue, the Neogene red beds of the Cheleken peninsula, contain approximately 2 ppm Cu in solution, ranging up to extremes of 15 ppm (Lebedev, 1972). Free thermal convection within the deep aquifer, involving recycling of pore fluids, would alleviate the high fluid requirement of single pass mechanisms and also allow for lower concentrations of ore constituents in the fluids, because with each convective pass the recycled fluids would have a chance to be recharged by renewed leaching of reservoir rocks. Consequently, free convection provides for a flushing mechanism that may collect essentially all leachable metal from an aquifer and make it available for precipitation in potentially giant ore deposits, provided a mechanism for focussed precipitation is in place. This is how we view the role of free convection in the formation of the Athabasca unconformity-type uranium deposits and we speculate that, by analogy, a similar process may have been active in the formation of the large stratiform copper deposits. In Proterozoic continental rocks which generally lack significant organic matter,

convective flushing may have been more efficient than in Phanaerozoic rocks containing numerous reducing traps, thus offering a potential for larger deposits.

The energy to initiate and drive convective flow in deeply buried sandstone may be provided by increased heatflow associated with periods of tectonic basin reactivation, considering that the formation of many of the stratiform and unconformity-related ore deposits under discussion appears to have been coincident with episodes of tensional tectonics and mafic magmatism during advanced stages of basin development (Sawkins, 1976; Annels, 1984; Lancelot et al., 1984).

With regard to the stratiform copper deposits, favorable factors for the establishment of free convection would have been (a) a stratigraphic development characterized by thick sections of fluviatile sandstone and conglomerate, containing few intercalations of marine shales; (b) a rift tectonic setting characterized by high heat flow (Perrodon and Masse, 1984); (c) episodes of tectonic basin reactivation and basaltic magmatism (Annels, 1984); (d) presence of basement highs on the floor of the red bed basin. By analogy with the Athabasca basin, the overlying marine shales may have constituted a low-permeability seal to the underlying red bed aquifer, conducive to the possible development of a geopressured zone.

The situation along the flanks of basement highs protruding into overlying red bed sandstones would be analogous to that of an inclined sandstone aquifer, for which Wood and Hewett (1982; 1984) have theoretically shown that unicellar convection could be induced by very low geothermal gradients. If the local thermal gradients are enhanced by conductivity contrasts between metamorphic rocks of the basement hills

and sandstones of the aquifer, more vigorous convection is to be expected. This suggests that the general association of stratiform copper deposits with basement highs may possibly be interpreted in terms of convective rolls on the flanks of such highs. The annular ring-shape of certain deposits (e.g. in the Chambishi South-East Prospect, Zambia; Annels and Simmonds, 1984; Fleischer, 1984) may reflect the operation of torroidal convective cells, as described by Wood and Hewett (1984), around basement highs. Patterns of metal distribution and mineral zonation, as well as the occurrence of barren, pyritic gaps.

For the Zambian Copper Belt we suggest that the stratiform copper ores and the local occurrences of mineralized veins in basement rocks where stratiform ores wedge out against basement highs, may both have formed by convection-mediated, diagenetic-hydrothermal mineralization at, repectively, the top and the base of the sandstone aquifer. A similar interpretation may apply to the Michigan Copper District, for which it has been argued that stratiform ore hosted by the Nonesuch Shale at the top of the Copper Harbour red bed aquifer and vein ore in the Portage Lake Volcanics at its base result from one process of epigenetic mineralization (Livnat et al., 1983; Nishioka et al., 1984). Although stable isotope studies (Livnat, 1983; Livnat et al., 1983) indicate a burial metamorphic, compactional origin for the mineralizing fluid, this does not argue against convective flushing of the Copper Harbour aquifer and convection-mediated, diagenetic-hydrothermal mineralization.

Viewed from our perspective, mineralization constitutes an inherent aspect of diagenesis and basin evolution, which opens the prospect of developing a unified geological model for a broad spectrum of ore de-

posits, whose unifying feature is an origin that is intimately tied to red bed diagenesis. At one end of the spectrum are unconformity-type uranium deposits; Gabon and Lodève-type stratiform uranium deposits, giant stratiform copper deposits and unconformity-type lead deposits, which formed during advanced stages of basin evolution. At the other end are syngenetic to early-diagenetic deposits including sandstone-type uranium and vanadium deposits, red bed copper deposits and certain types of sandstone lead deposits. The Olympic Dam Cu-U-Au deposit in South-Australia, which displays similarities to stratiform copper deposits (Roberts and Hudson, 1983), may belong to this same family of deposits, as may the Shinkolobwe and associated uranium deposits in Shaba, Zaire. For the latter deposits, Nongo-Kashisha (1975) and Craig and Vaughan (1979) have shown that the U-Ni-Co-Cu-Fe-V-Mo-Au-Pt-Pb-Se-S mineralization is similar to Kamoto-type copper deposits with respect to stratigraphic and structural setting, stratiform character, metal assemblage and paragenetic sequence.

We speculate that the enigmatic, mineralized pipes of Tslumeb and Kombat in Namibia (Söhnge, 1964; Haynes, 1984; Lombaard et al., in pres) and of Kipushi, Zaire (Intiomale and Oosterbosch, 1974) also belong to this category. In a preliminary report, Hughes et al. (1984) classify these deposits as belonging to, what they term, Tsumeb-type mineralization, which also include the deposits of M'Passa (Congo Republic) and of Lombe and Kengere in Zaire. Tsumeb-type mineralization is associated with unconformities that coincide with the transition from marine carbonate to sandstone molasse sedimentation. The mineralization is located in solution-collapse structures in carbonate sequences, adjacent to basement highs. Preliminary lead isotopic data of Hughes et al. (1984)

suggest that epigenetic ore in the pipes and stratiform copper mineralization in overlying sandstones may have a common origin related to major basin dewatering events that permeated the entire stratigraphic sequence. Tsumeb-type solution pipes may mark locations where formation waters flowed through impermeable carbonate horizons. This must have taken place, possibly intermittently, during advanced stages of basin evolution and under conditions of high-grade diagenesis, because fluid inclusion evidence for the Tsumeb deposit indicates temperatures of approximately 250°C (Haynes, 1984).

The apparently time-bound character of unconformity-type uranium deposits, stratiform copper deposits and stratiform uranium deposits such as at Lodève and in Gabon (Jacobsen, 1975; Robertson et al., 1978; Toens and Andrews-Speed, 1984) may suggest that crustal and tectonic conditions allowing the process of basin evolution, red bed diagenesis and ore formation to preceed to completion have rarely been met in the Earth's history. Unconformity-type uranium deposits are known primarily form the Middle-Proterozoic. Stratiform copper deposits appear in the geologic record in early Proterozoic times but have their main development during the Middle-Proterozoic at ~1000 Ma, and a second major phase during the Permian. These were both periods when the land areas on earth were assembled in gigantic supercontinents, creating conditions that were conducive to the formation of extensive, intracratonic basins, widespread fluvatile terrestrial sedimentation and red bed formation.

FIGURES

AND

PLATES

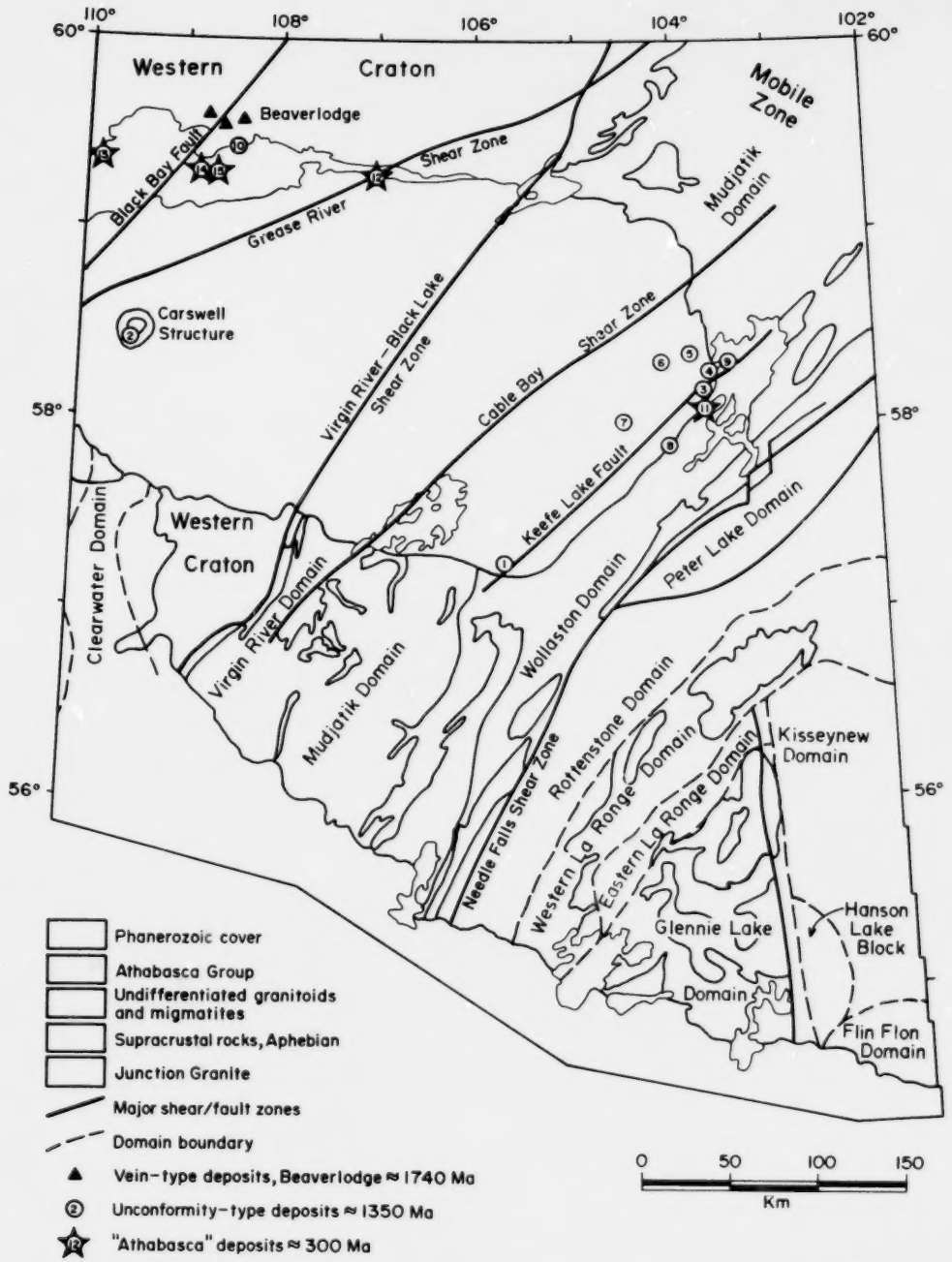


Figure 1: Schematic geological map of northern Saskatchewan (modified from Lewry and Sibbald, 1978), showing major crustal subdivisions and location of uranium deposits. (1) Key Lake; (2) Cluff Lake; (3) Rabbit Lake; (4) Collins Bay; (5) McClean Lake; (6) Midwest Lake; (7) Cigar Lake; (8) West Bear Lake; (9) Eagle Point; (10) Nicholson; (11) Raven and Horseshoe; (12) Fond-du-Lac; (13) Maurice Bay; (14) Johnson Island; (15) Stewart Island.

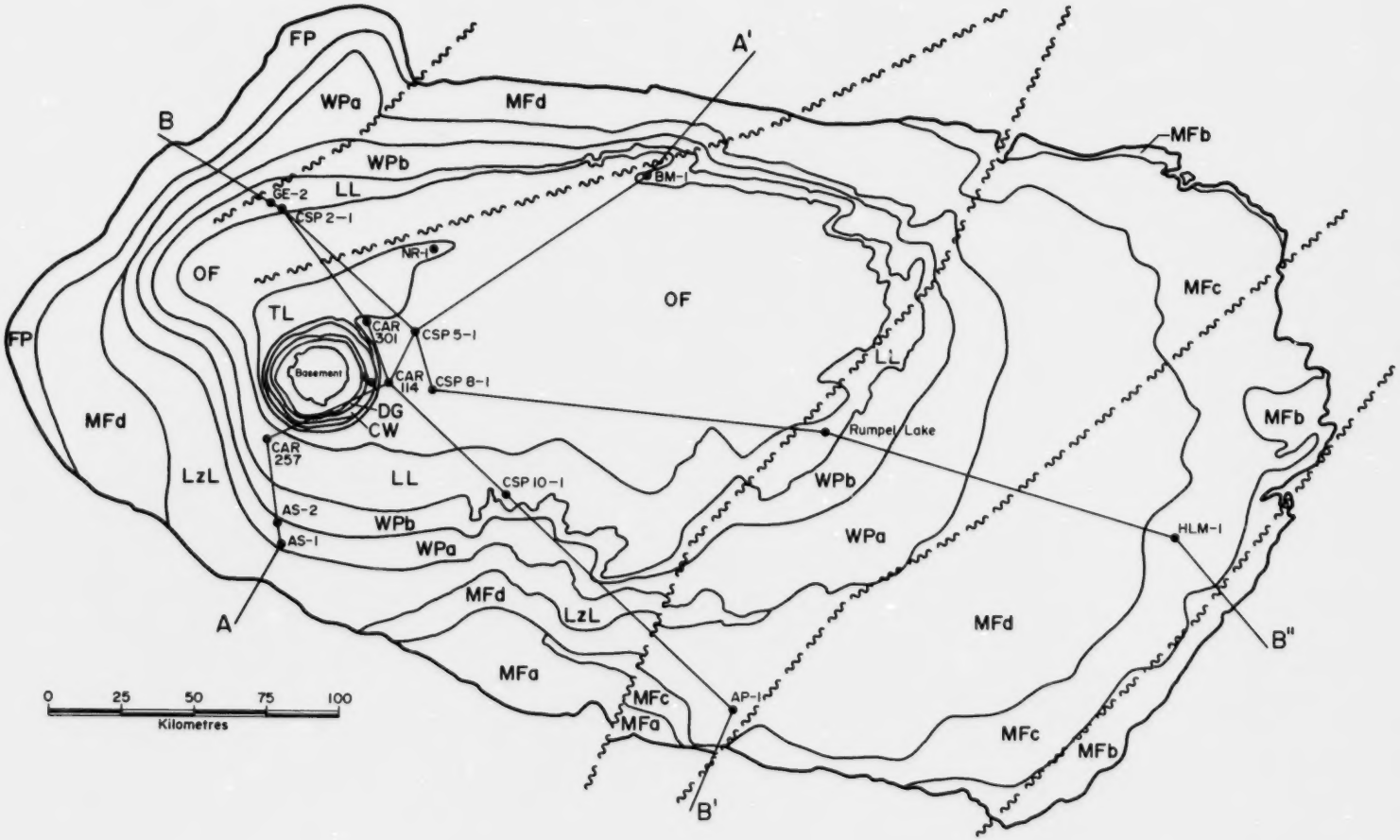


Figure 2: Geology of the Athabasca basin (modified from Ramaekers, 1979; 1980; Wilson, 1985; this study) with location of drill holes and major cross-section lines. Formation boundaries to the south and west of the Carswell Structure are approximate only. Stratigraphic symbols after Ramaekers, 1979; 1980: MF - Manitou Falls Formation; LzL-Lazenby Lake Formation; FP-Fair Point Formation; WP-Wolverine Point Formation; LL-Locker Lake Formation; OF-Otherside Formation; TL-Tuma Lake Formation; DG-Douglas Formation; CW-Carswell Formation. Legend as in Figure 3.

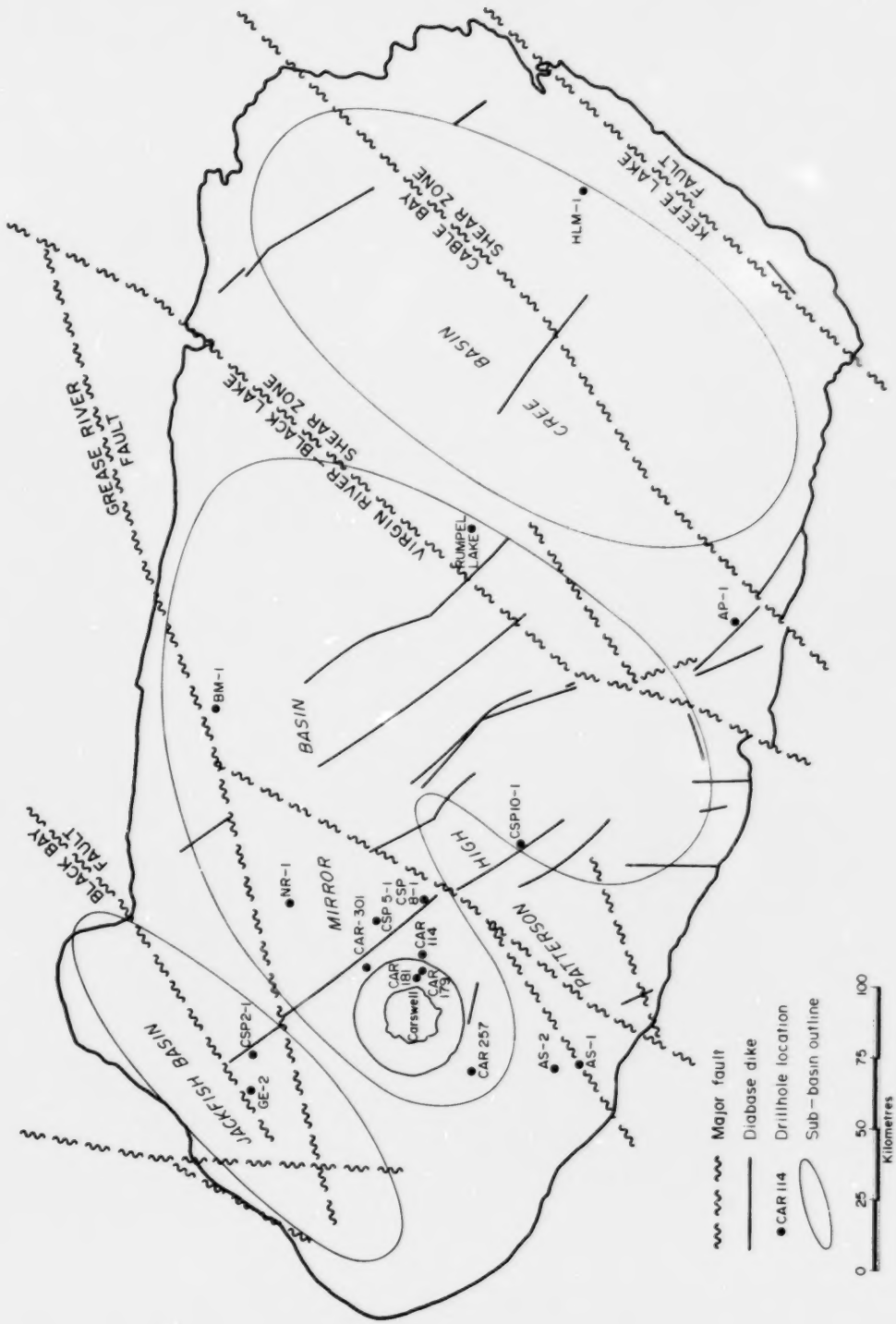


Figure 3: Sub-basins of the Athabasca basin (modified from Ramaekers 1979; 1980; Hobson and McAulay, 1969).

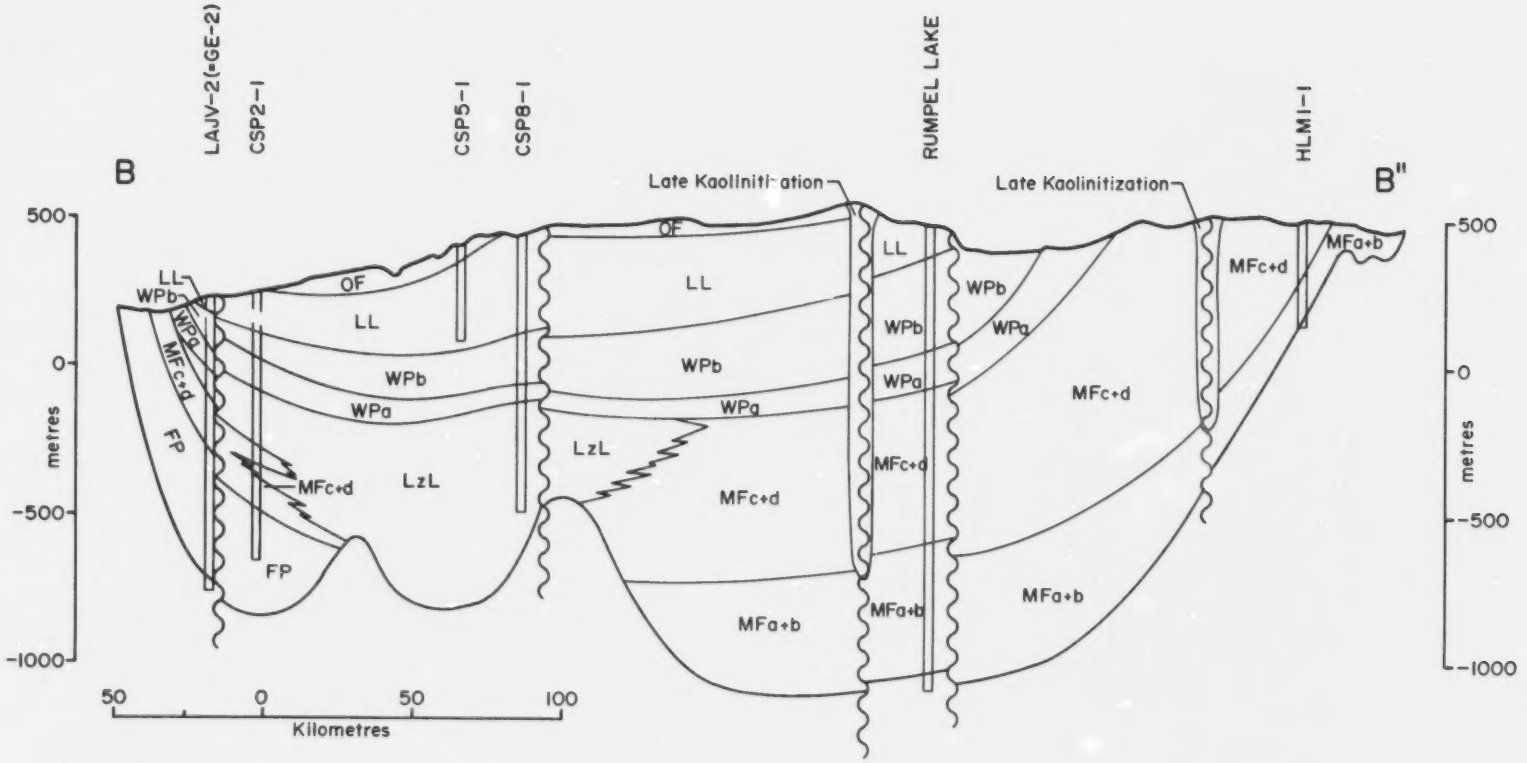


Figure 4: Interpreted clay mineral stratigraphy in section B-B"; late fault-controlled kaolinitization schematically indicated only; legend as in Figure 2.

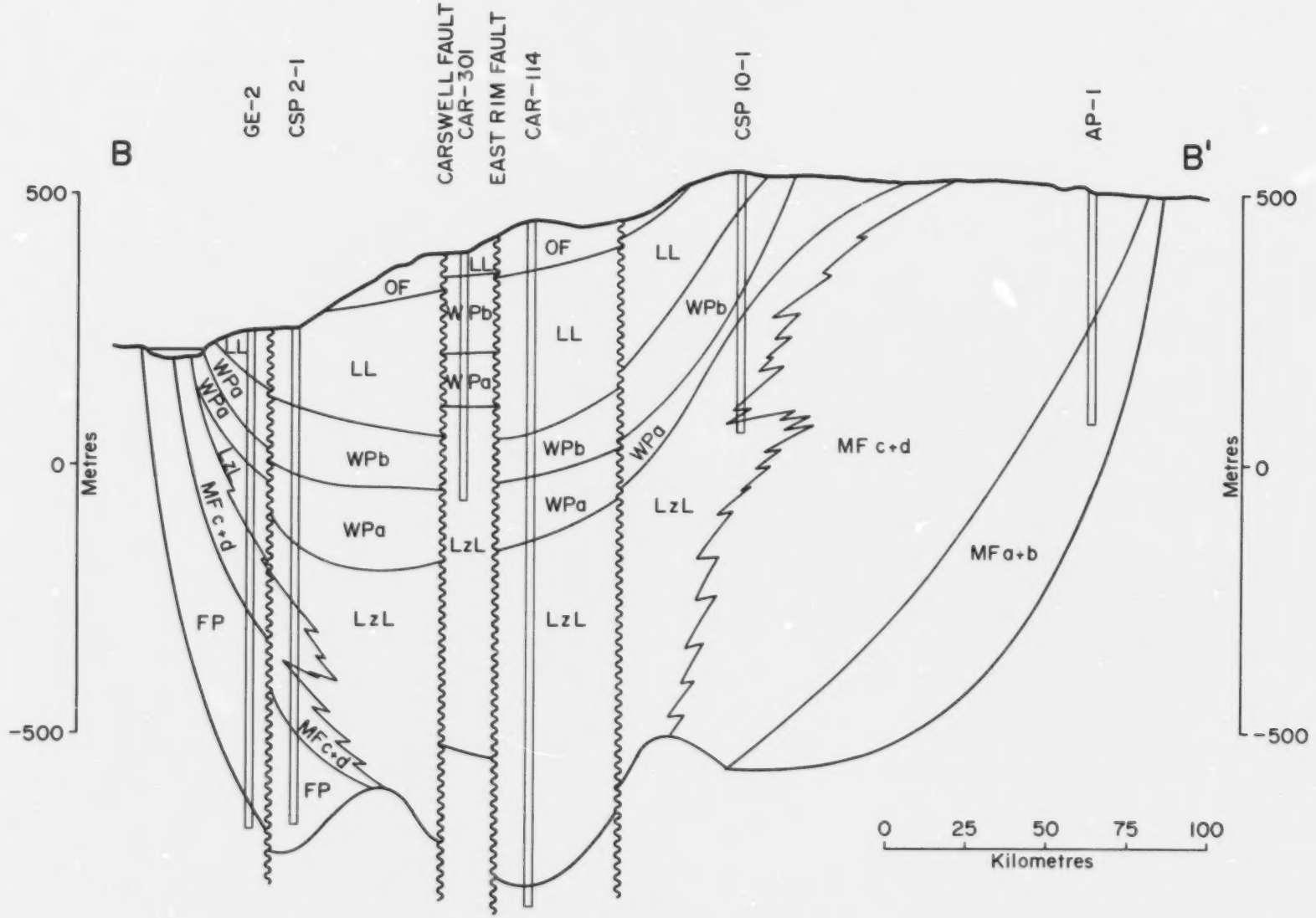


Figure 5: Interpreted clay mineral stratigraphy in section B-B'; legend as in Figure 2.

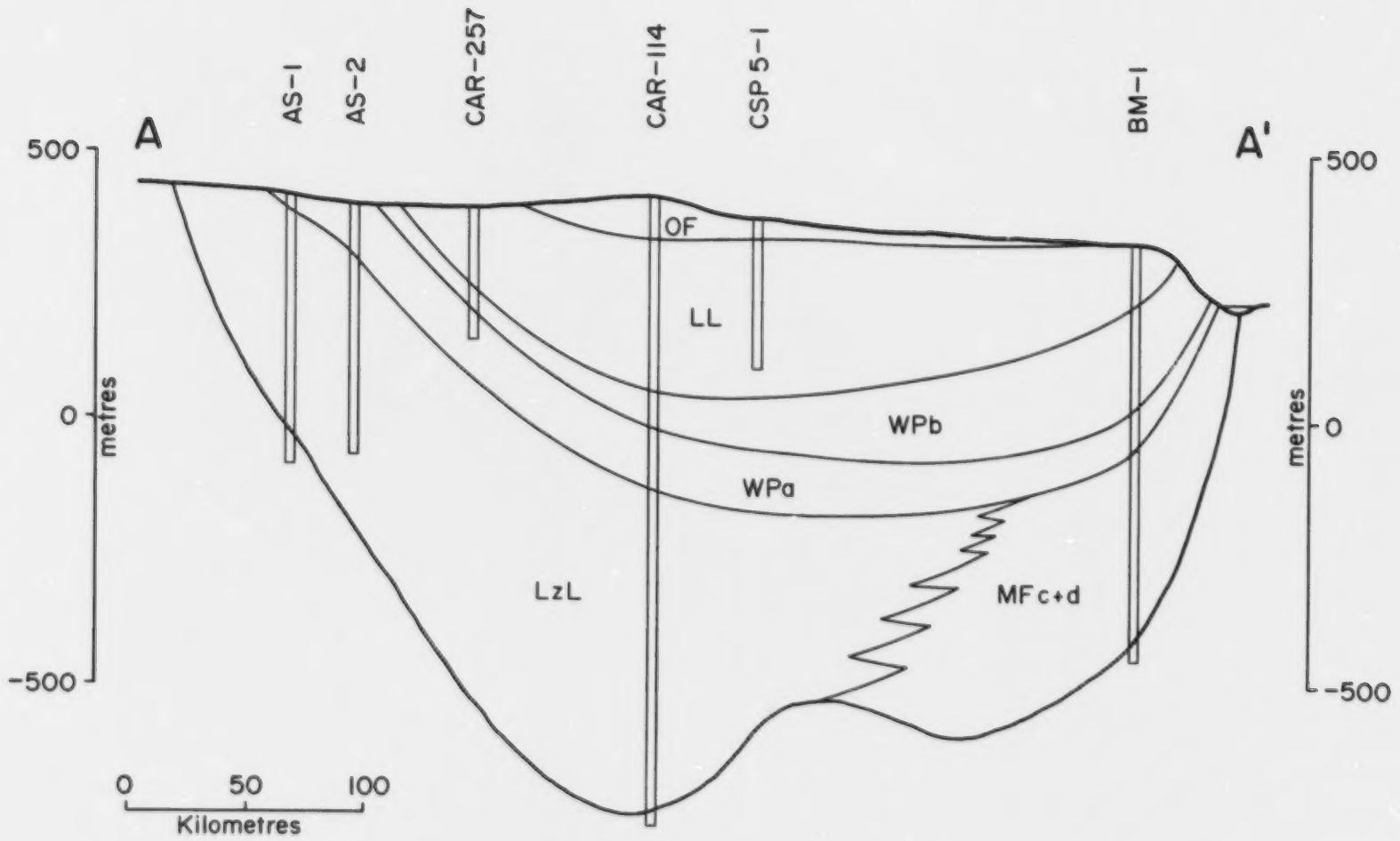
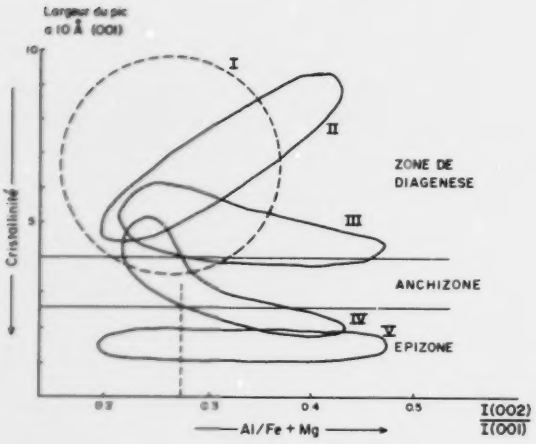
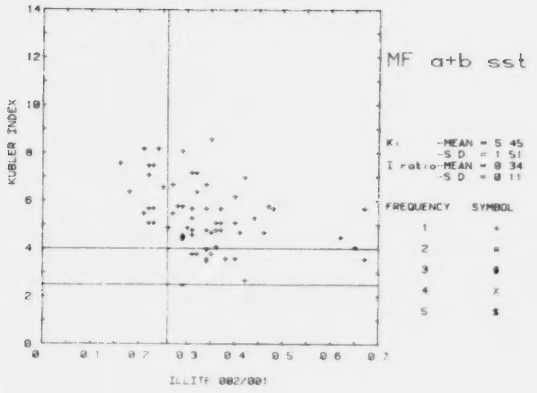
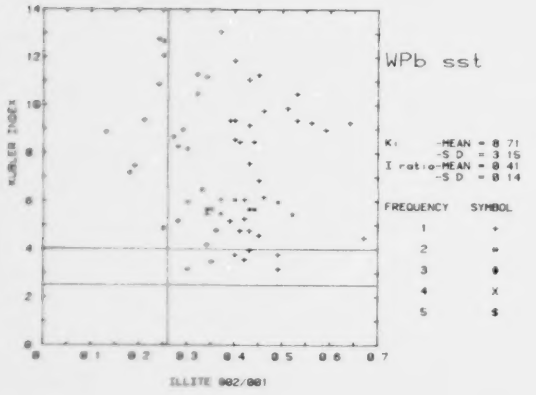
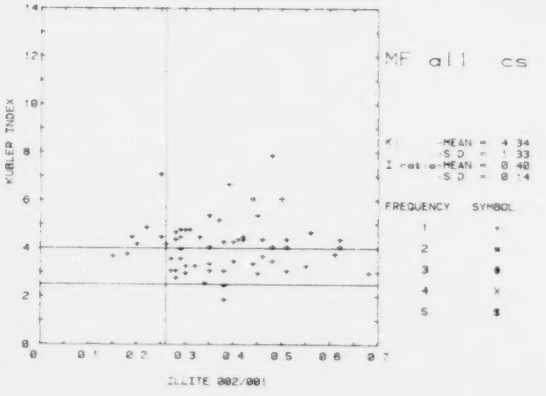


Figure 6: Interpreted clay mineral stratigraphy in section A-A'; legend as in Figure 2.









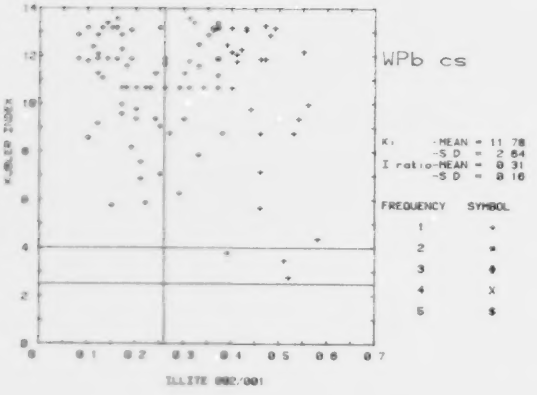


Figure 7: Esquevin diagrams for Athabasca illites. A - general dia. (after Dunoyer de Segonzac, 1968) showing fields of (1) incipient diagenesis; (2) low-grade diagenesis; (3) high-grade diagenesis; (4) anchizone; (5) epizone or low-grade metamorphisms; B - diagram for MFa+b sandstones; C - diagram for WPb sandstones; D- diagram for MF clay/silt seams; E - diagr. for WPb clay/silt seams.

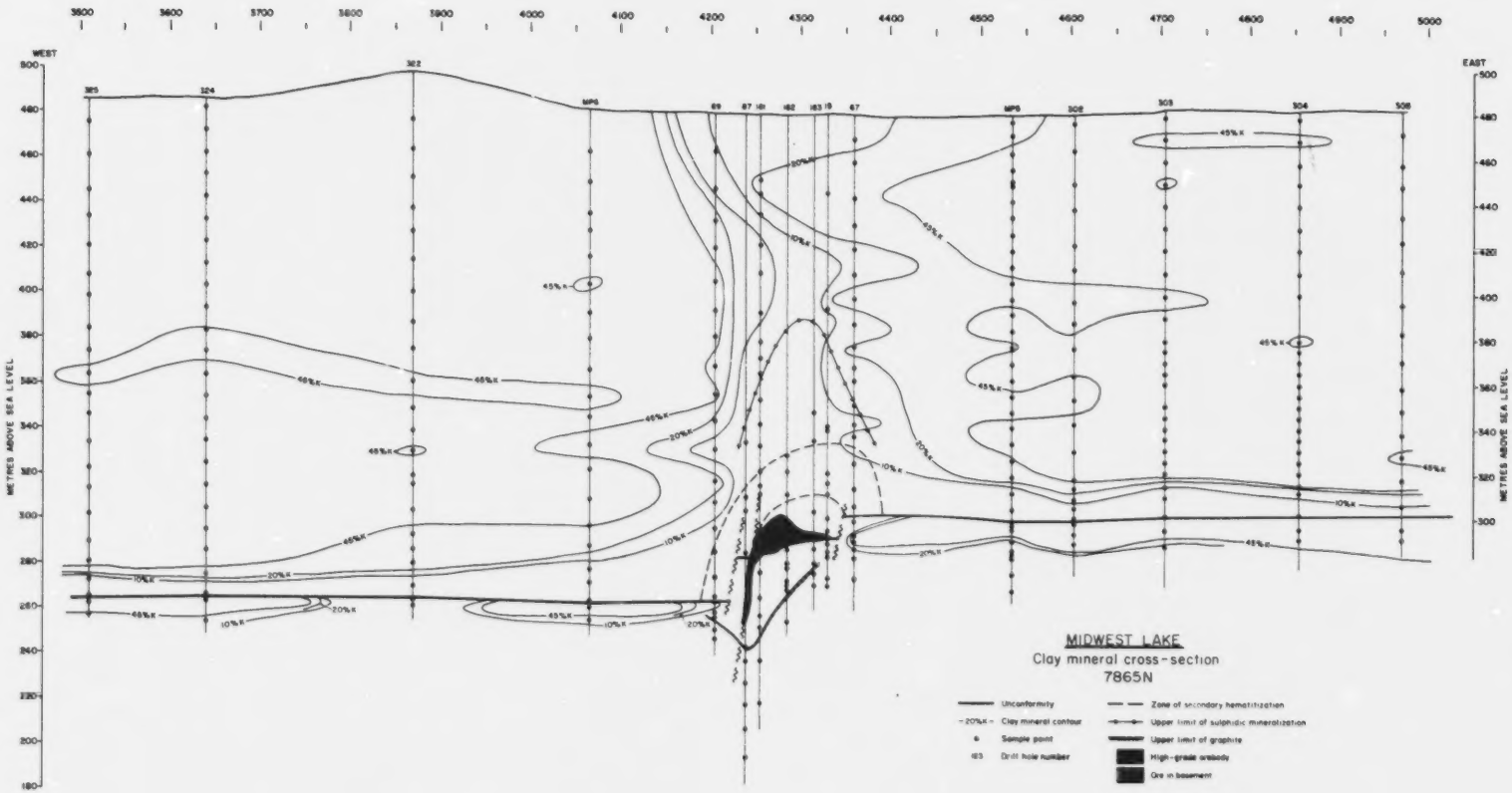


Figure 8: Midwest deposit: clay mineral alteration haloe in section 7865N through main ore body; contours denote relative percentages of Kaolinite.

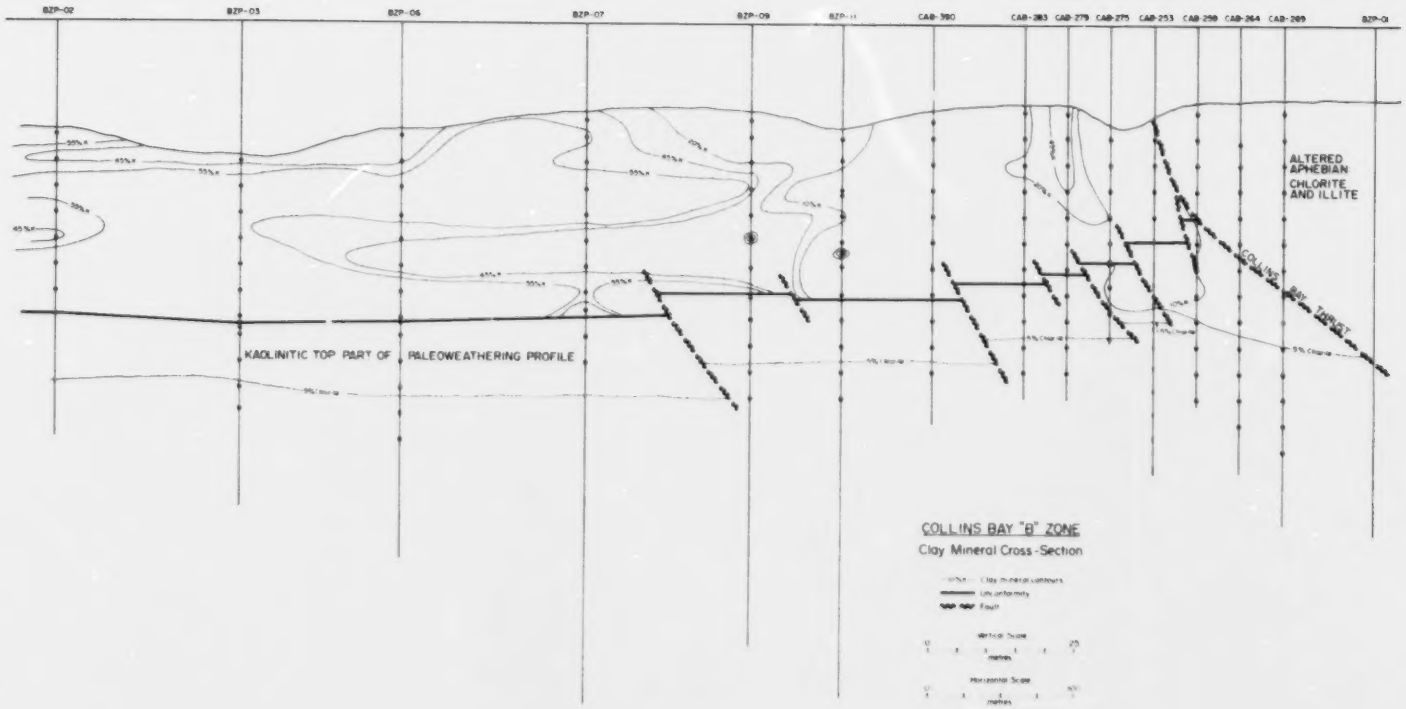


Figure 9: Collins Bay B deposit: clay mineral alteration haloe in cross section through northern part of ore body; contours denote relative percentages of kaolinite.

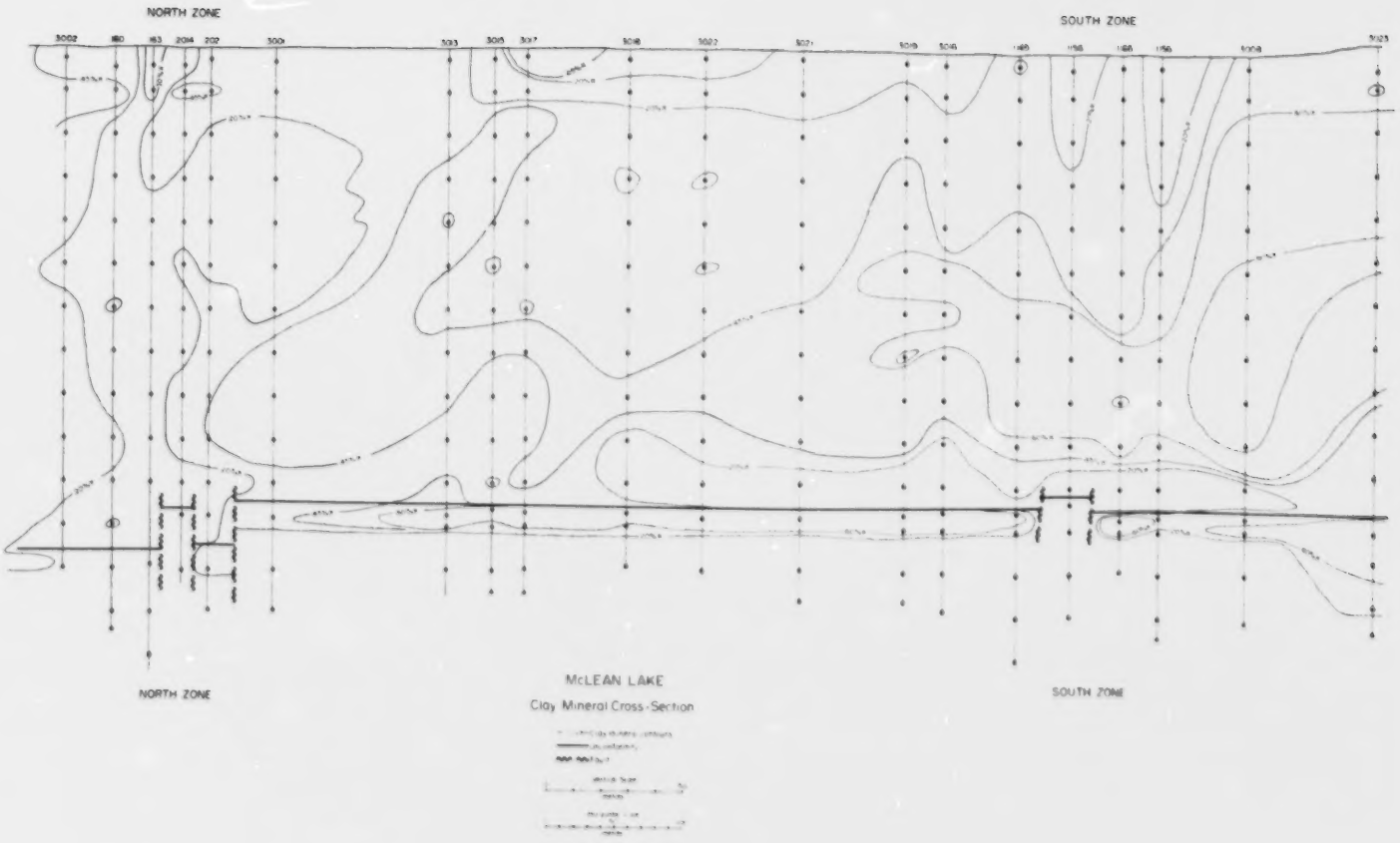


Figure 10: McClean Lake deposits: clay mineral alteration haloe in a central cross-section through northern and southern ore bodies; contours denote relative percentages of kaolinite.

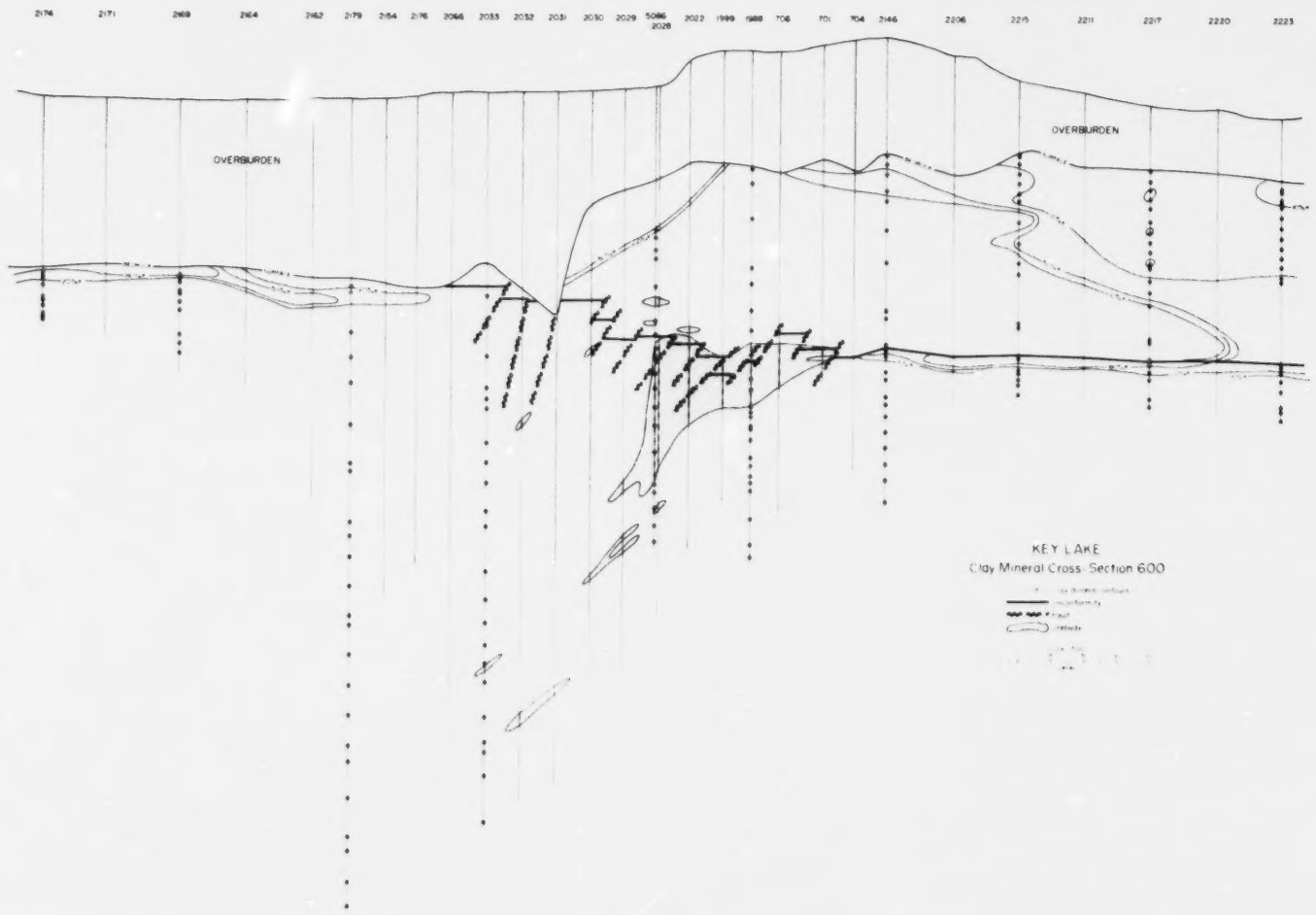
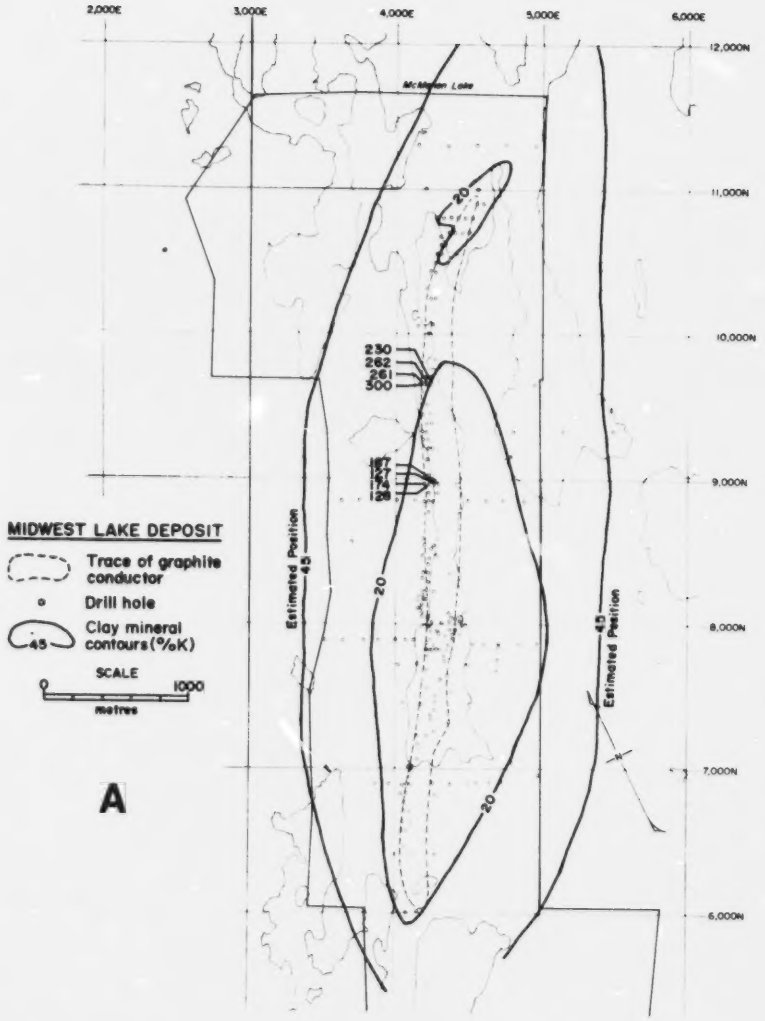
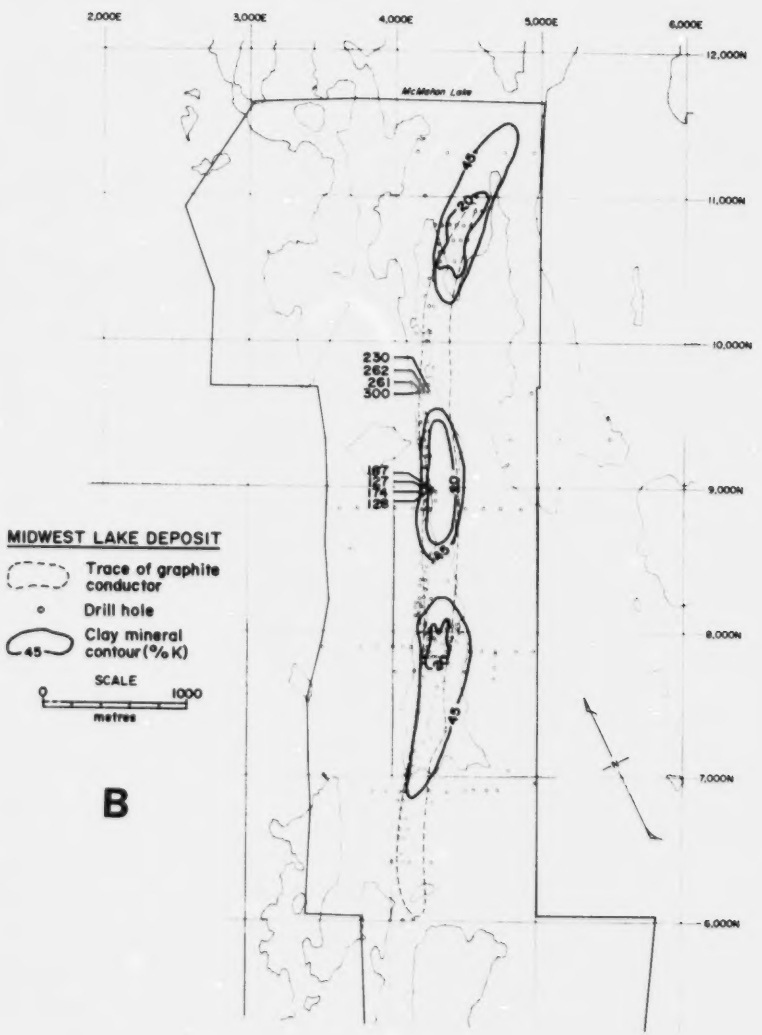


Figure 11: Key Lake: clay mineral alteration haloe in section 11 600 E through Deilmann ore body; contours delineate relative percentages of kaolinite.


2,000E
3,000E
4,000E
5,000E
6,000E
12,000N
11,000N
10,000N
9,000N
8,000N
7,000N
6,000N
McMahon Lake
20
230
262
261
300
167
127
74
28
Estimated Position
45
20
MIDWEST LAKE DEPOSIT
Trace of graphite conductor
Drill hole
Clay mineral contours (%K)
45
SCALE
0
1000
metres
A


2,000E
3,000E
4,000E
5,000E
6,000E
12,000N
11,000N
10,000N
9,000N
8,000N
7,000N
6,000N
McMahon Lake
45
20
230
262
261
300
167
127
74
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MIDWEST LAKE DEPOSIT
Trace of graphite conductor
Drill hole
Clay mineral contour (%K)
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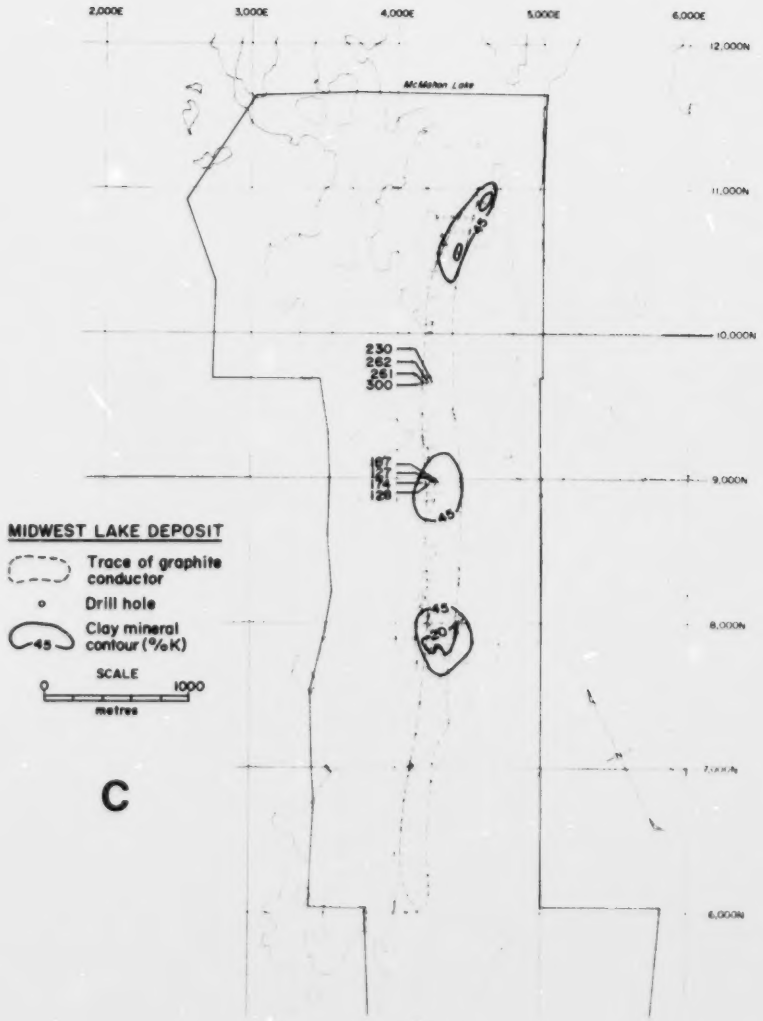

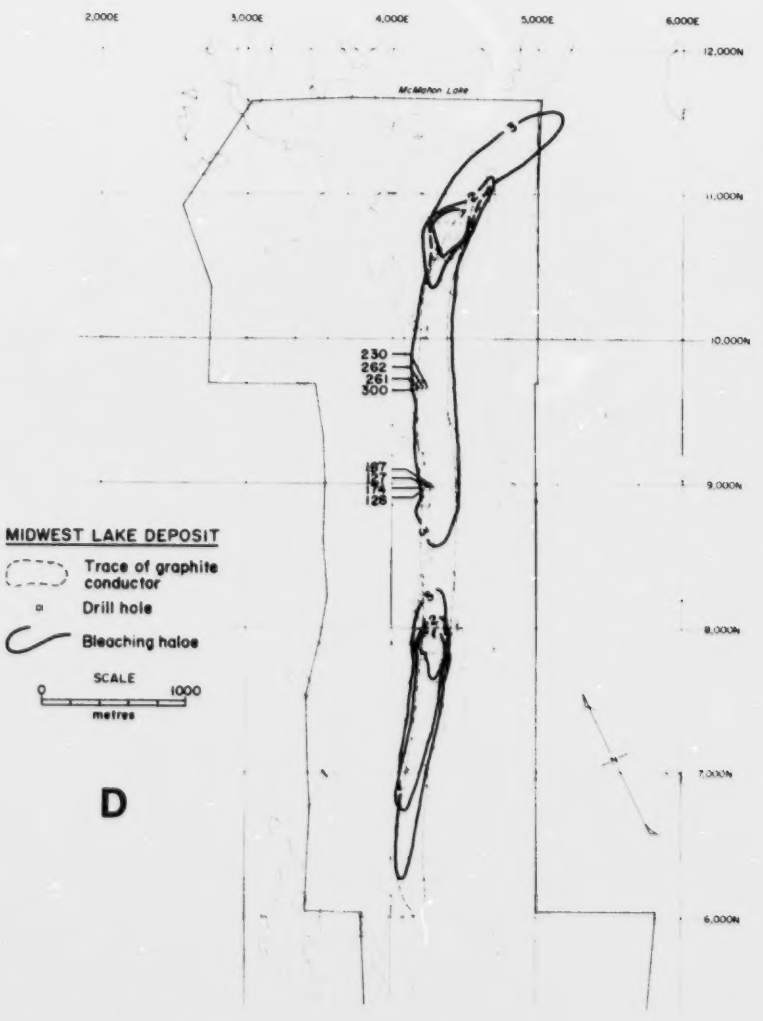

Figure 12: Midwest deposit: A,B,C - plan view of clay mineral alteration haloe; A: haloe in sandstone at unconformity; B: at 100 m above unconformity; C: at top bedrock, 200 m above unconformity. D - bleaching haloe in plan view; 1 - at top bedrock; 2 - 100 m above unconformity; 3 - at unconformity (top basement).

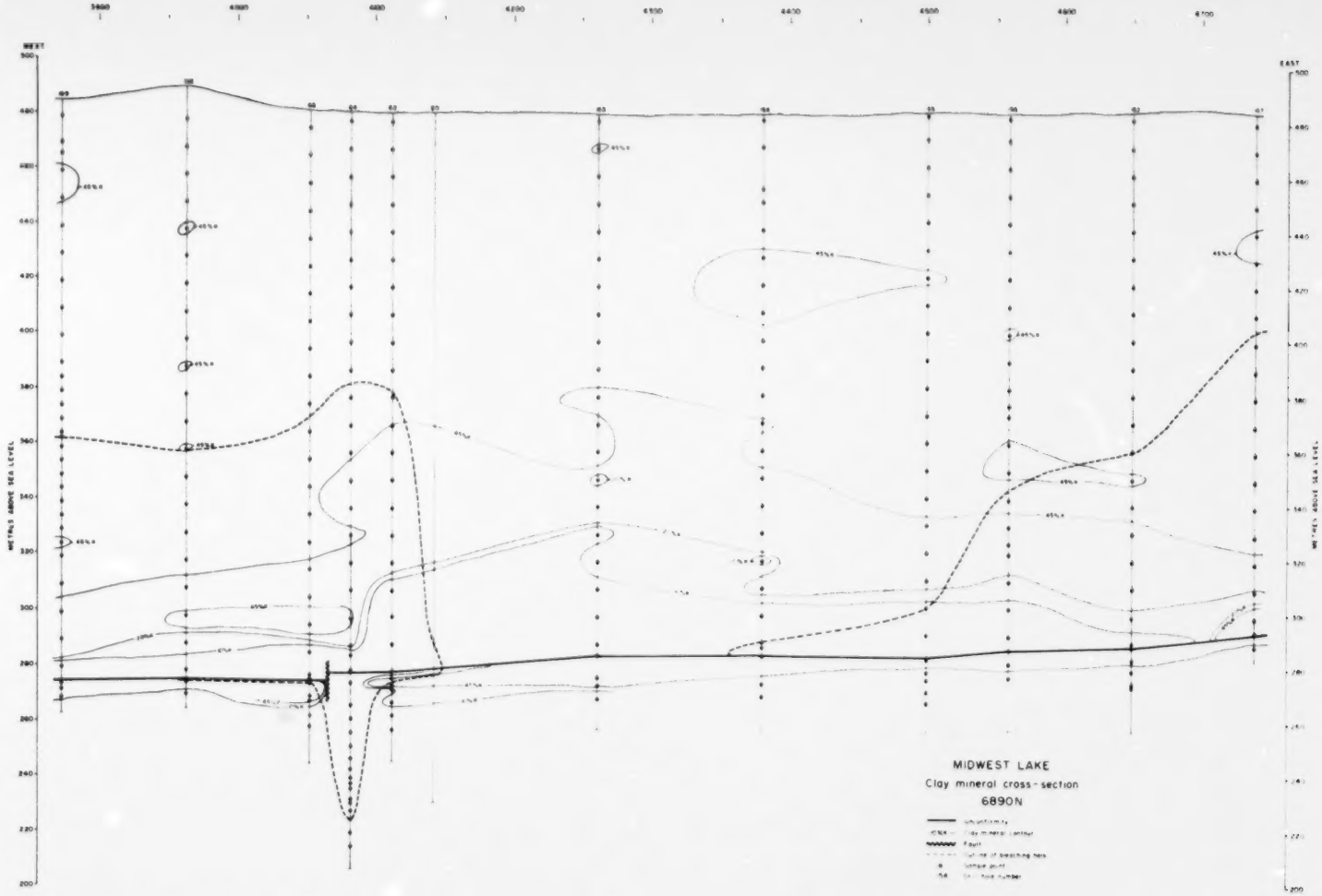

Figure 13: Midwest deposit: haloes of clay mineral alteration and bleaching in section 6890 N located near the southern termination of mineralized structure. Note that bleaching haloe cross-cuts clay contours, suggesting lateral flow from right to left.

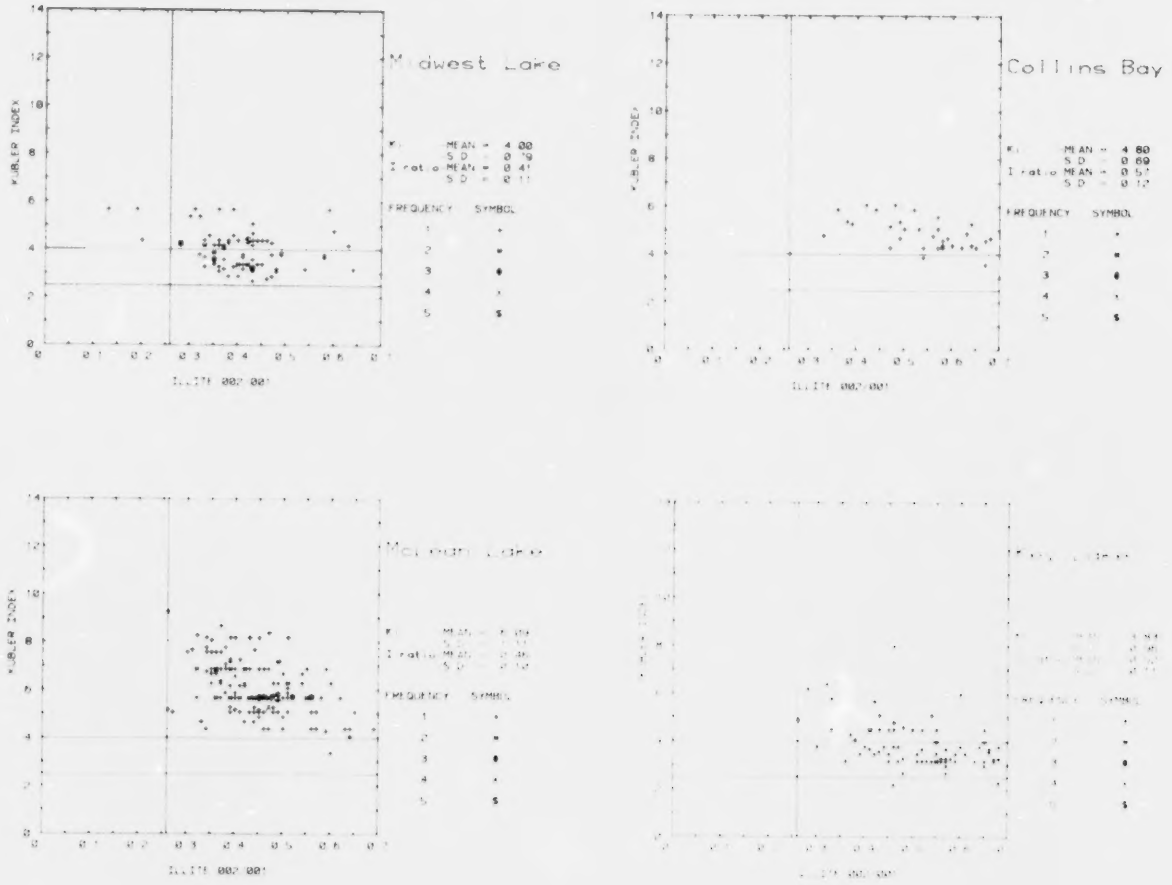

Figure 14: Esquevin diagrams for illites of alteration haloes at Midwest; Collins Bay B; McClean Lake; and Key Lake, (anomalously illitic sandstone).

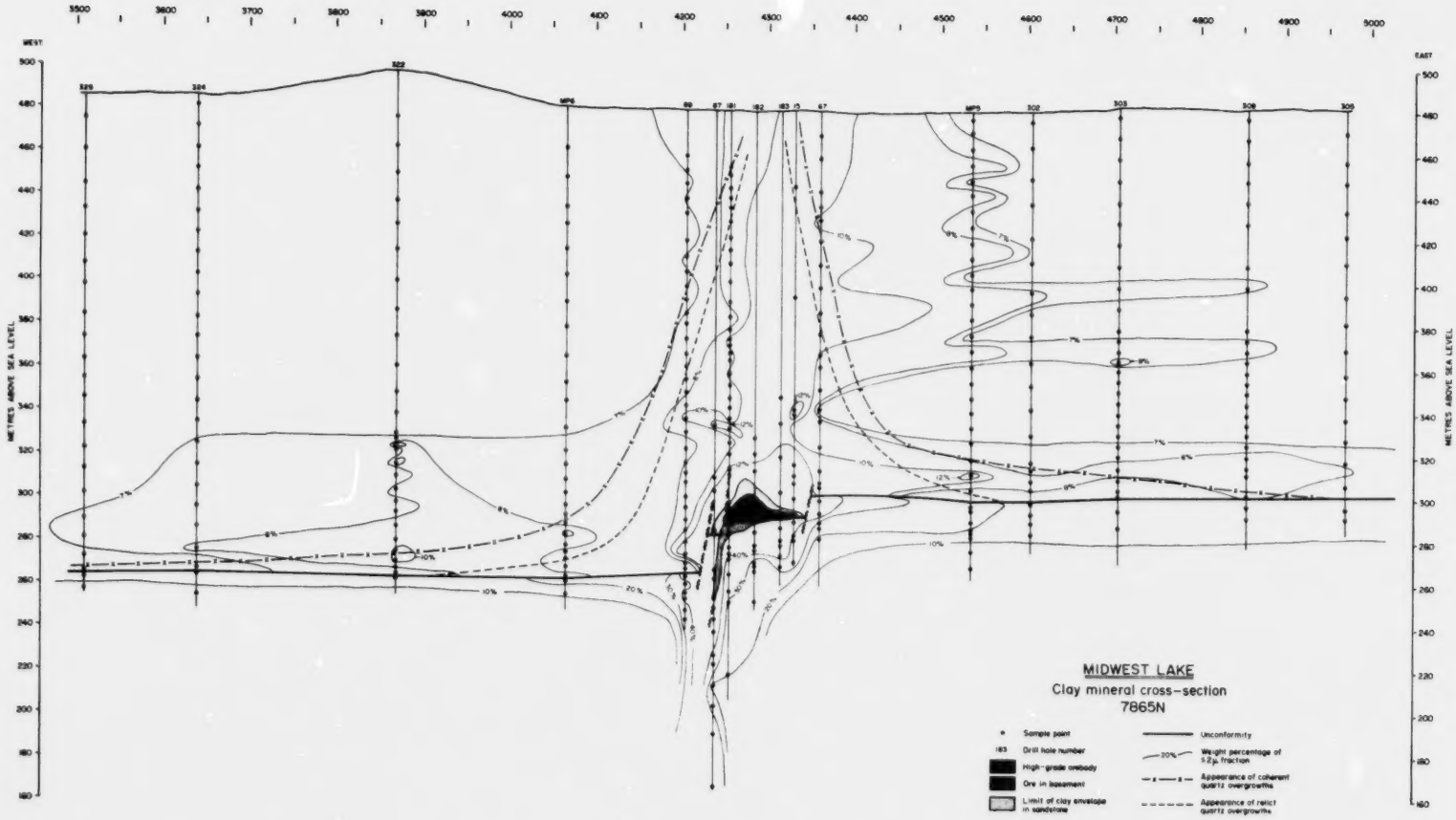

Figure 15: Midwest deposit: quartz corrosion and residual clay enrichment in section 7865 N; contours denote weight percentages of ≤2 μm fraction; heavy dashed lines delineate disappearance of quartz overgrowths.

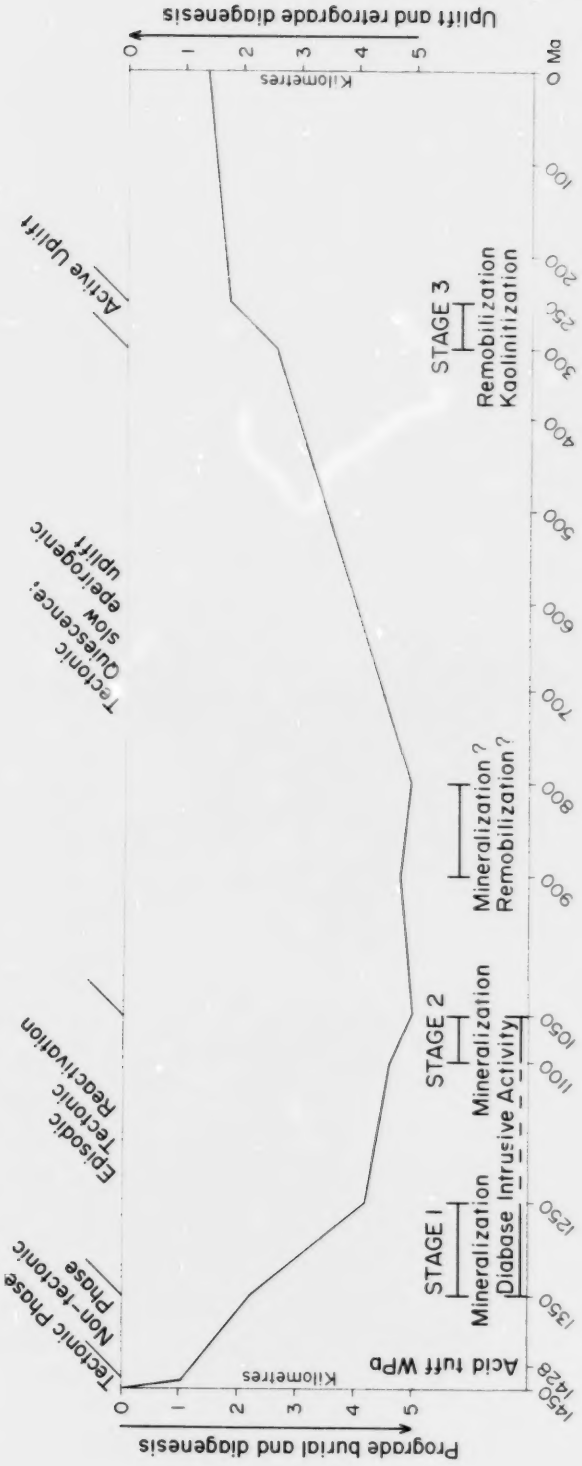

Figure 16: Schematic representation of geological evolution of the Athabasca basin.

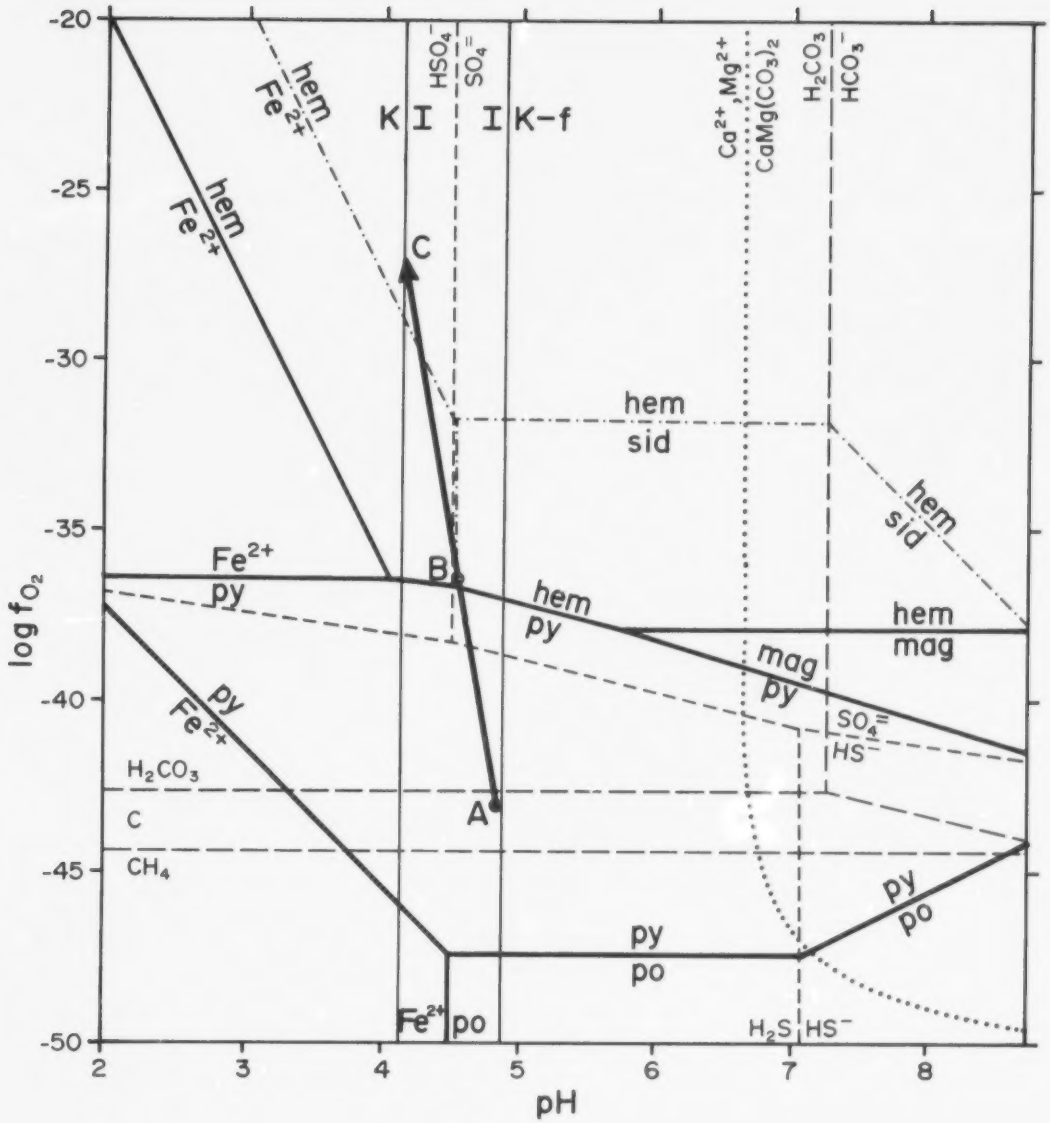

Figure 17: Log fO_2 - pH diagram constructed to illustrate conditions during first and second stages of mineralization; solid lines: first stage of mineralization; point-dashed lines: second stage; T = 200°C; $a^{\Sigma S} = 10^{-2}$; $a^{\Sigma C} = 10^{-1}$ (second stage: 10); $a^{Fe2+} = 10^{-5}$ (second stage: 10^{-7}); $a^{K+} = 2.5 \times 10^{-1}$; $a^{Ca++} = a^{Mg++} = 10^{-6}$. py-pyrite; po-pyrrhotite; hem-hematite; mag-magnetite; sid-siderite; K-kaolinite; I-illite; Kf - potassium feldspar.

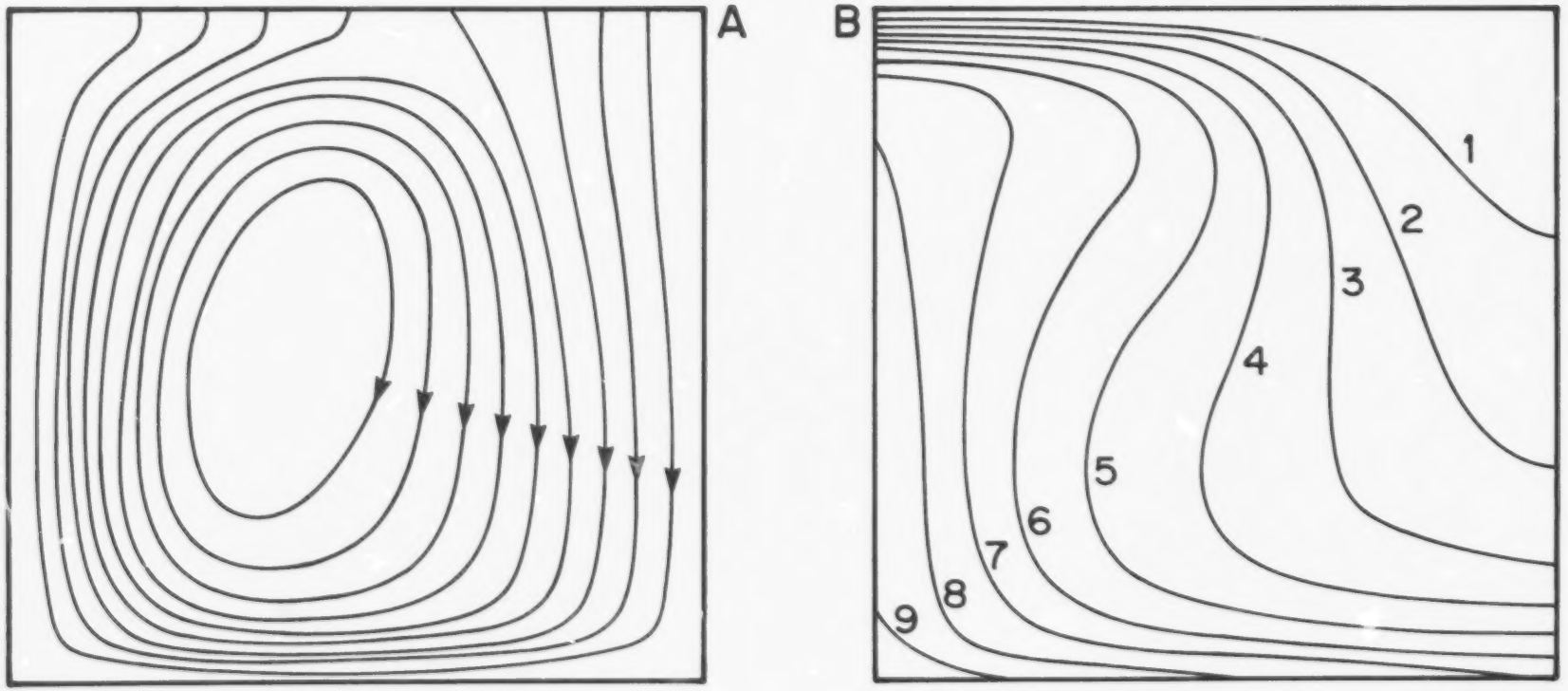

Figure 18: Stream lines (A) and isotherms (B) for sub-seafloor convection at mid-ocean ridge (after Parmentier and Spooner, 1978).

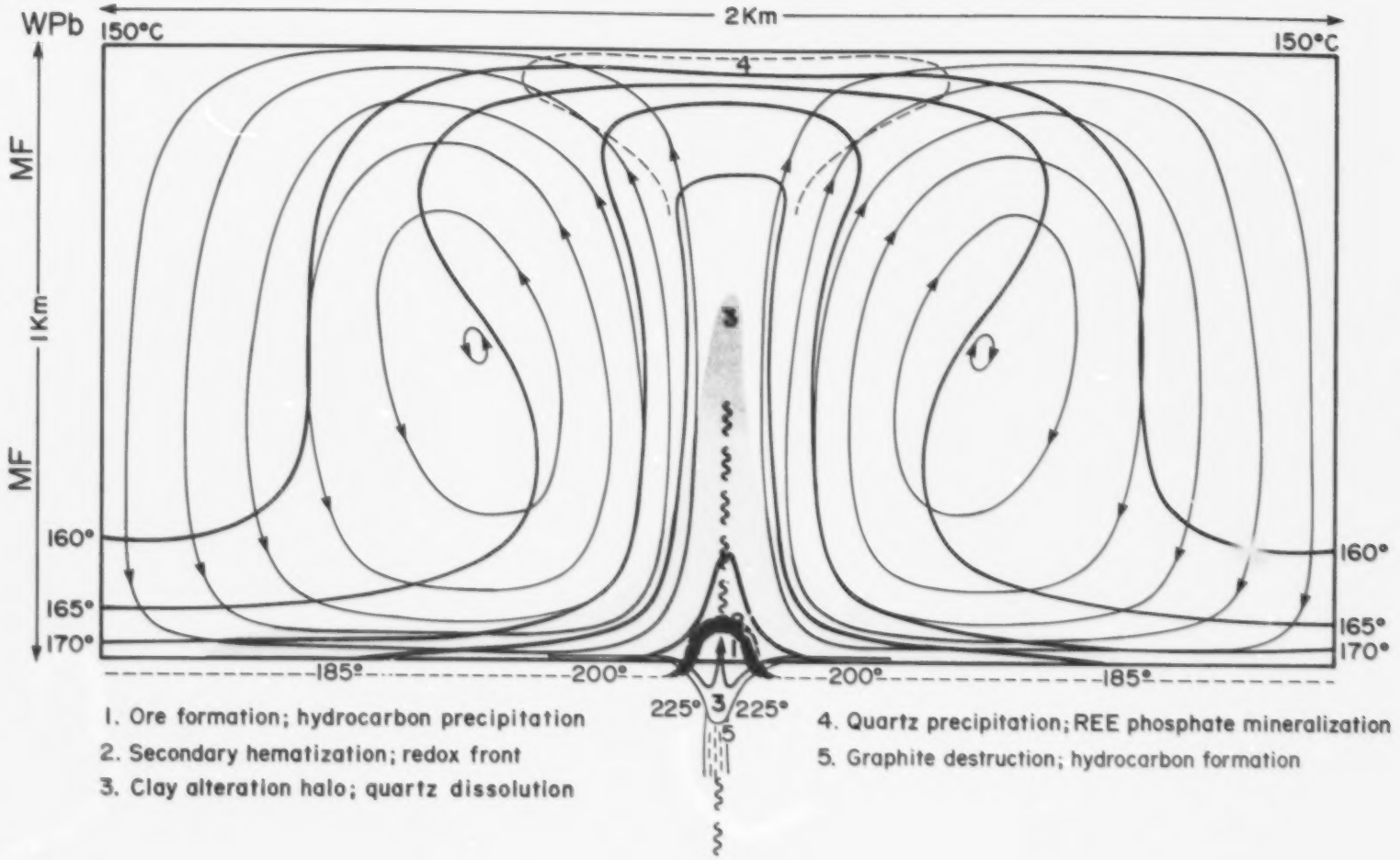

Figure 19: Schematic representation of diagenetic-hydrothermal convective system as inferred for formation of unconformity-type uranium deposits in Athabasca basin; configuration of isotherms and flow lines adapted from Parmentier and Spooner, (1978), and from quartz dissolution contours in Figure 15.

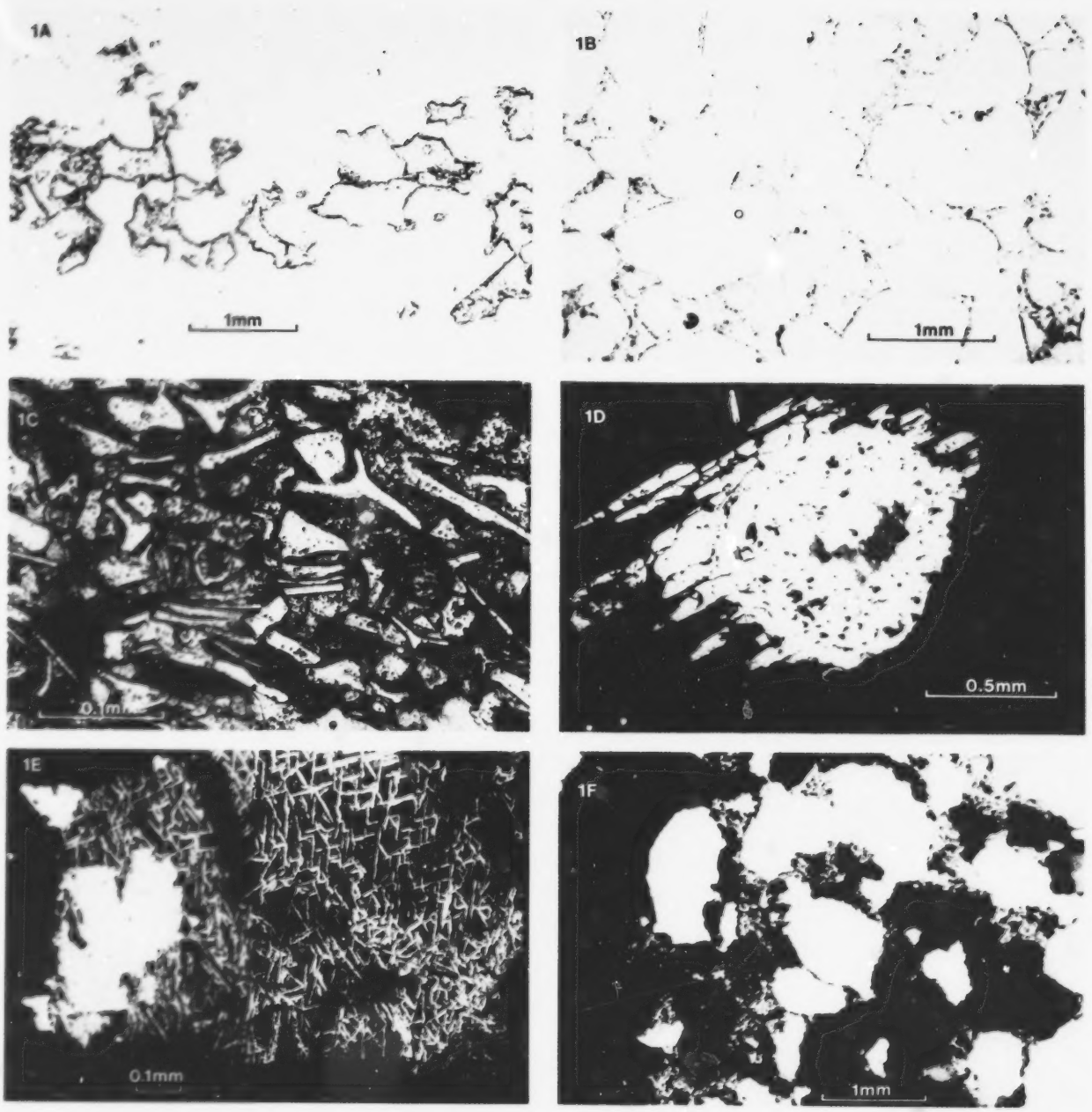

Plate 1 A - Poikiloblasts of diaspore enclosing detrital quartz.

B - Silica-cemented sandstone of Lazenby Lake Formation (LzL). Note absence of hematite pigment at grain contacts; predominance of point contacts.

C - Felsic tuff fragment in upper Wolverine Point Formation (WPb). Shard textures are preserved.

D - Hematite-rutile pseudomorphous after ilmenite and containing dark core of goethite-rutile pseudomorph. Authigenic specular hematite around detrital grain.

E - Chalcopyrite-siderite-rutile pseudomorphous after ilmenite; white: chalcopyrite; dark grey:siderite; medium grey:rutile lamellae.

F - Corroded quartz grains characteristic of clay alteration haloes; matrix consists of illite.

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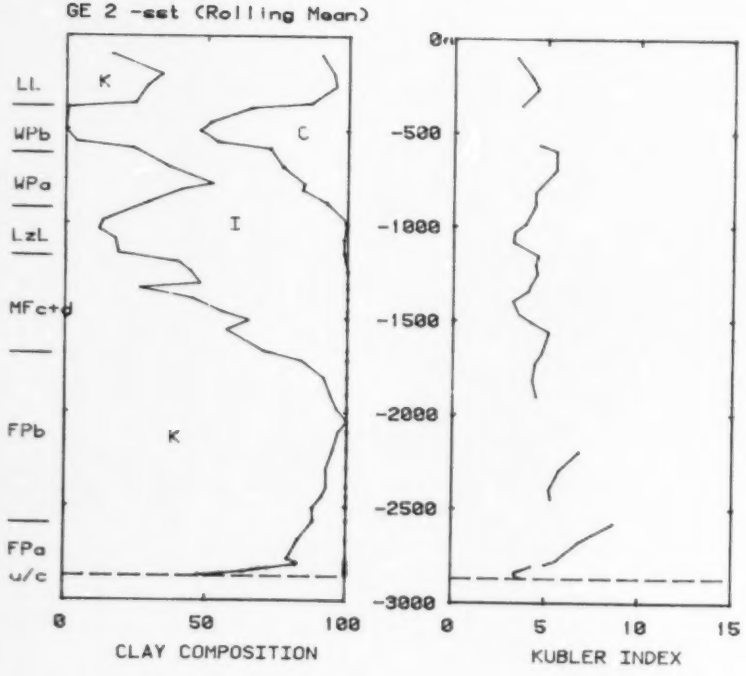
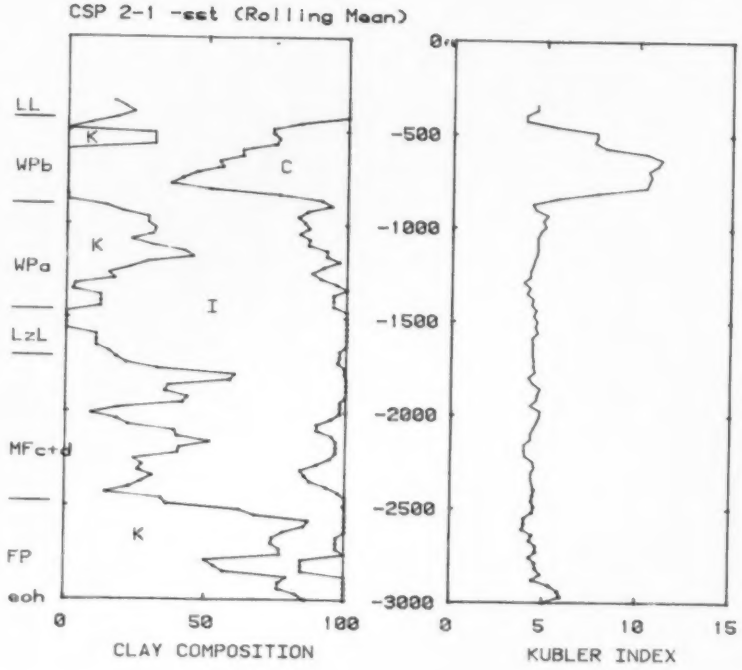
APPENDIX A





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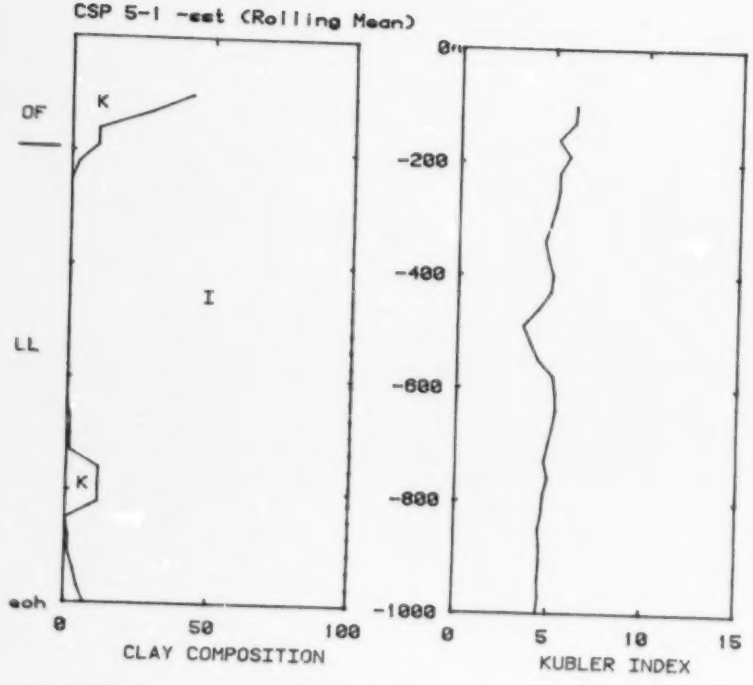

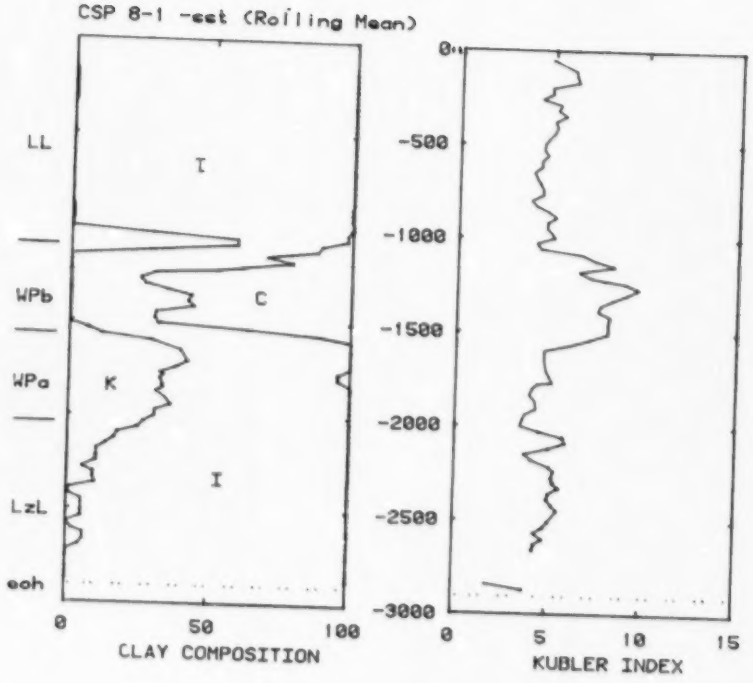

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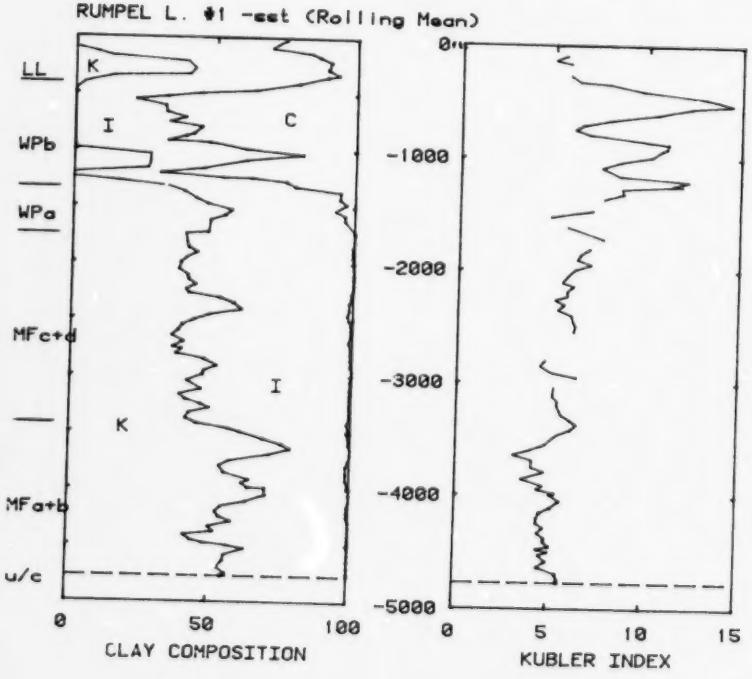

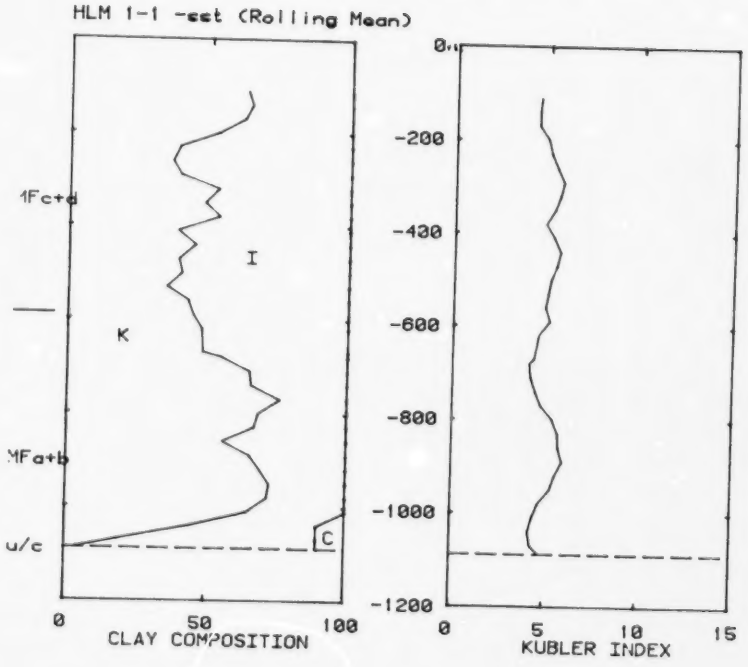

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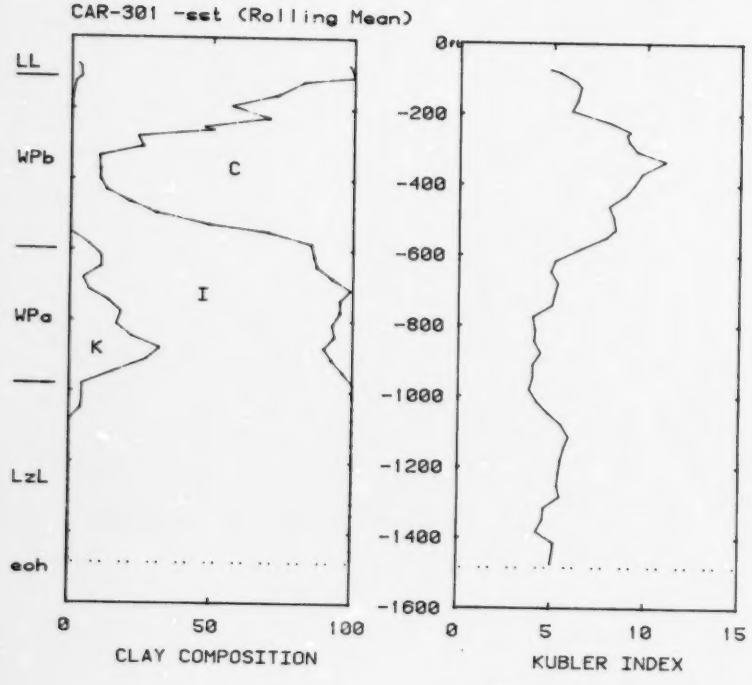

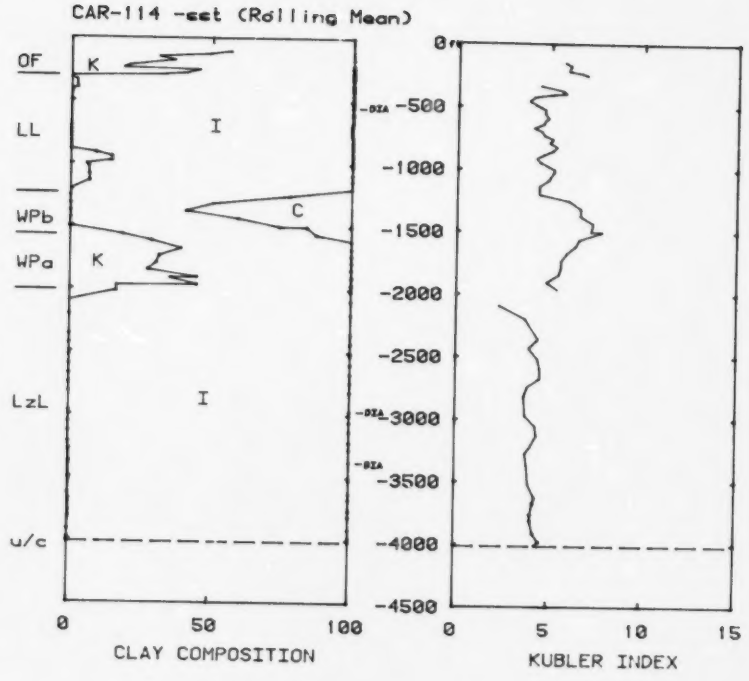

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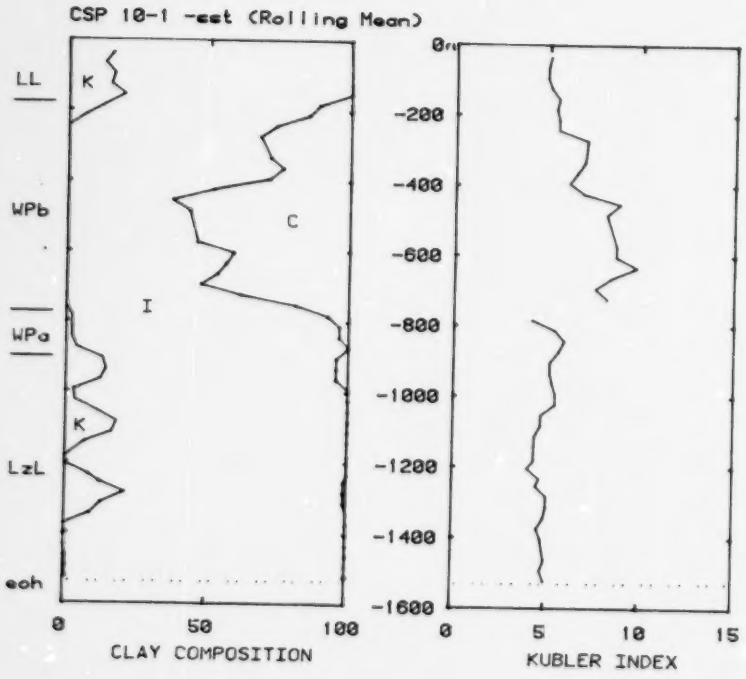

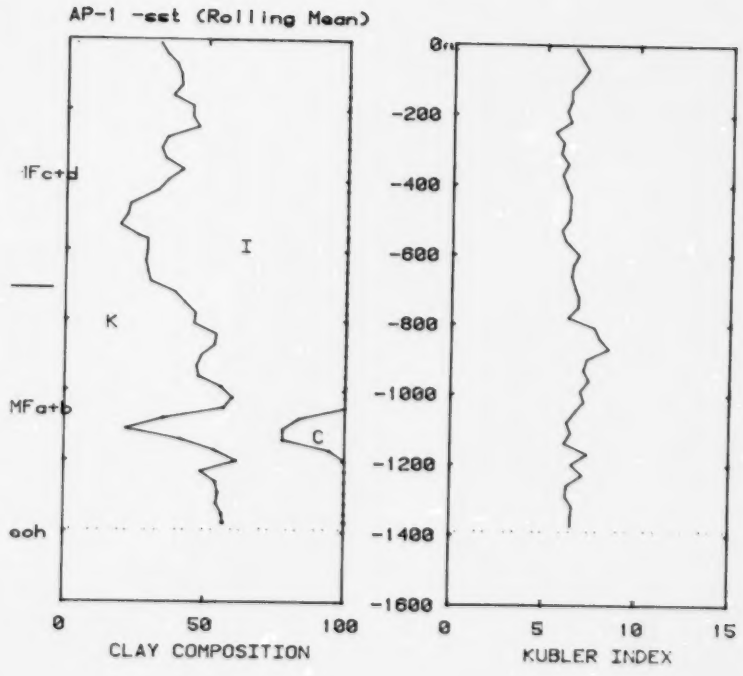

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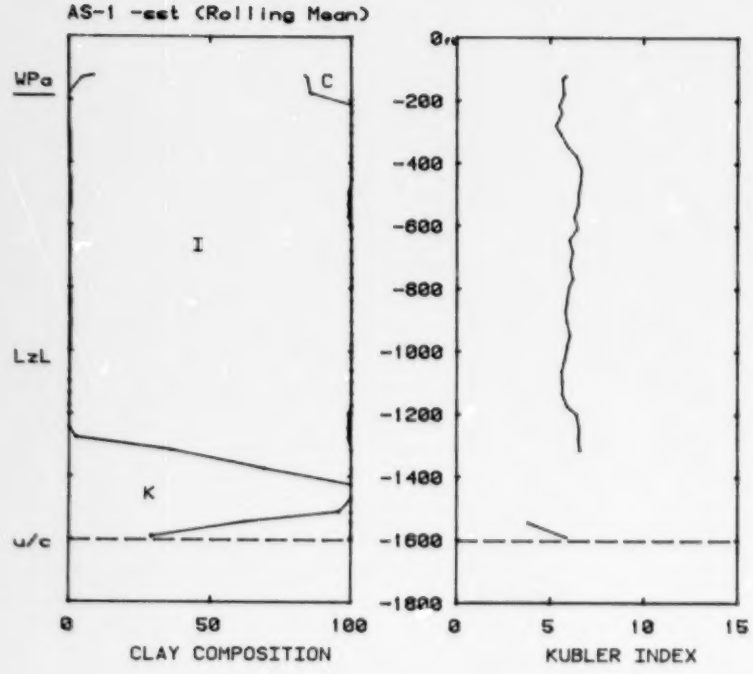

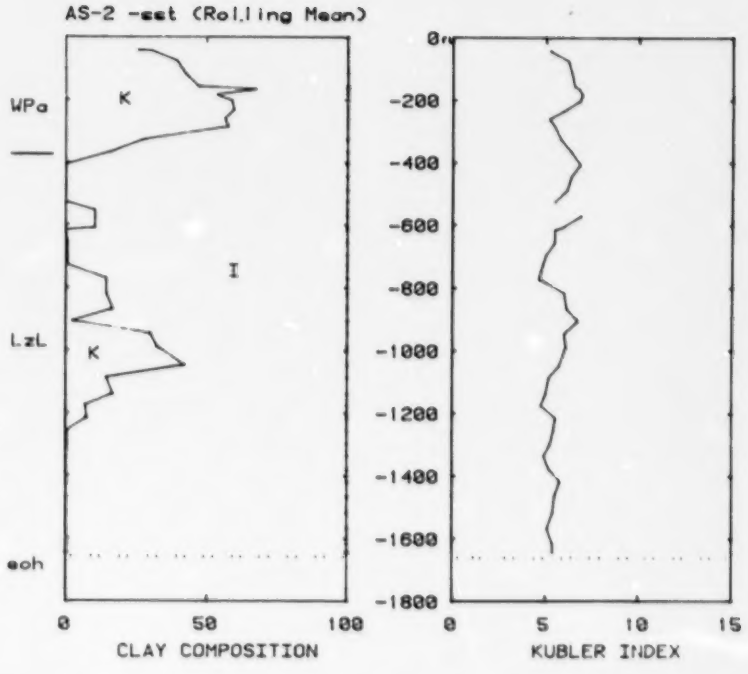

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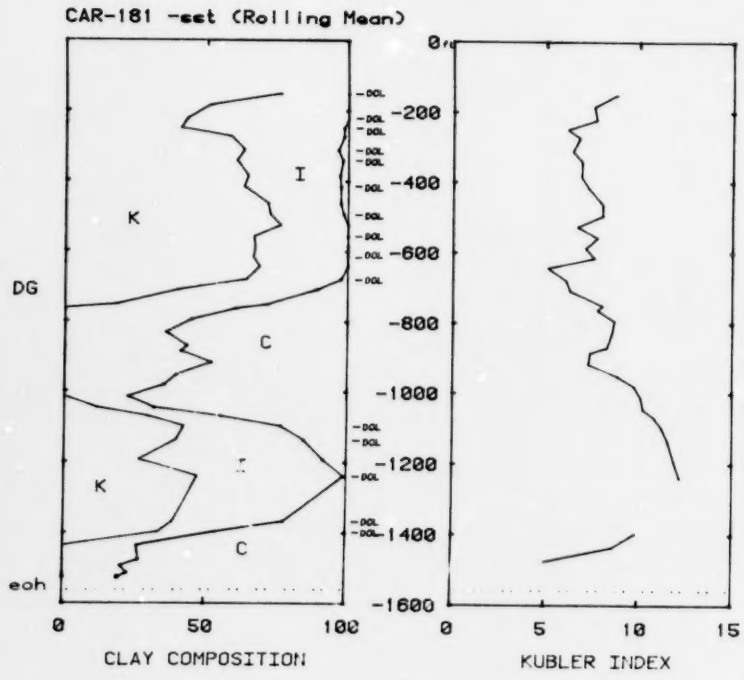

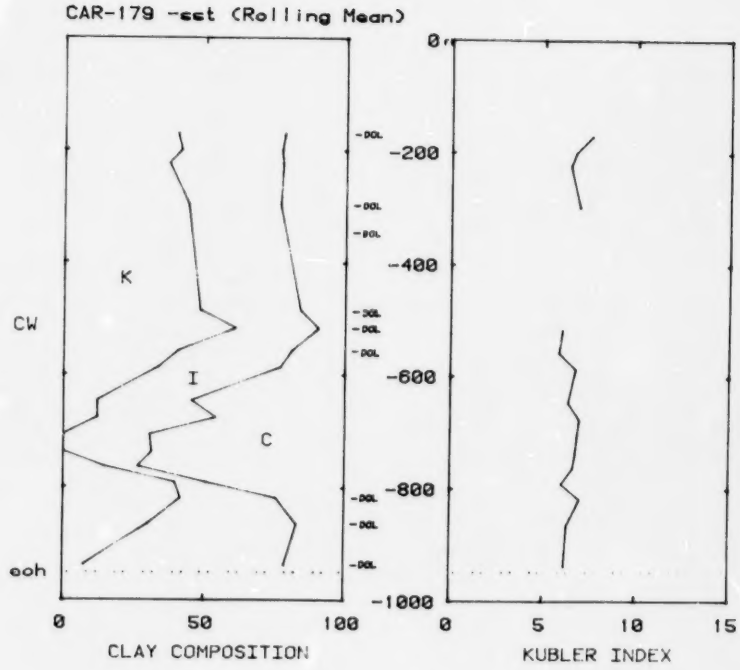

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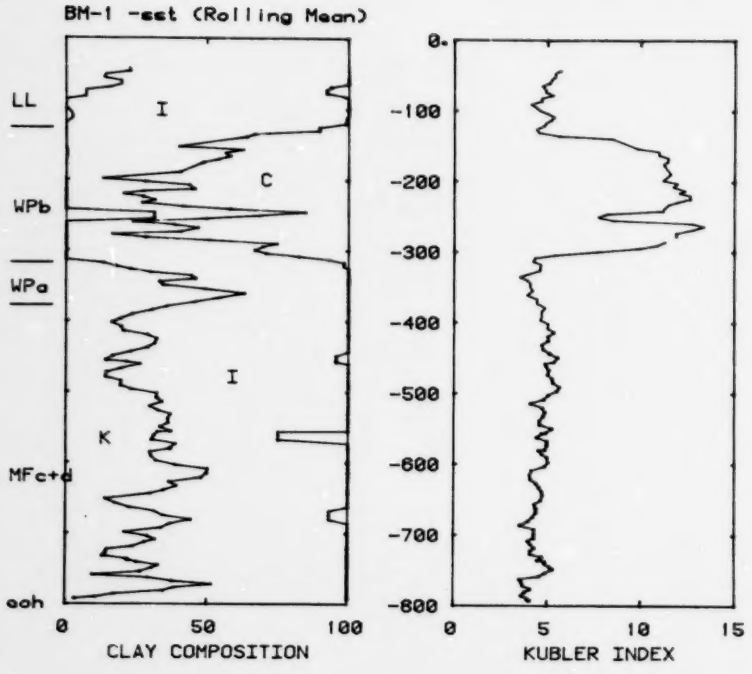

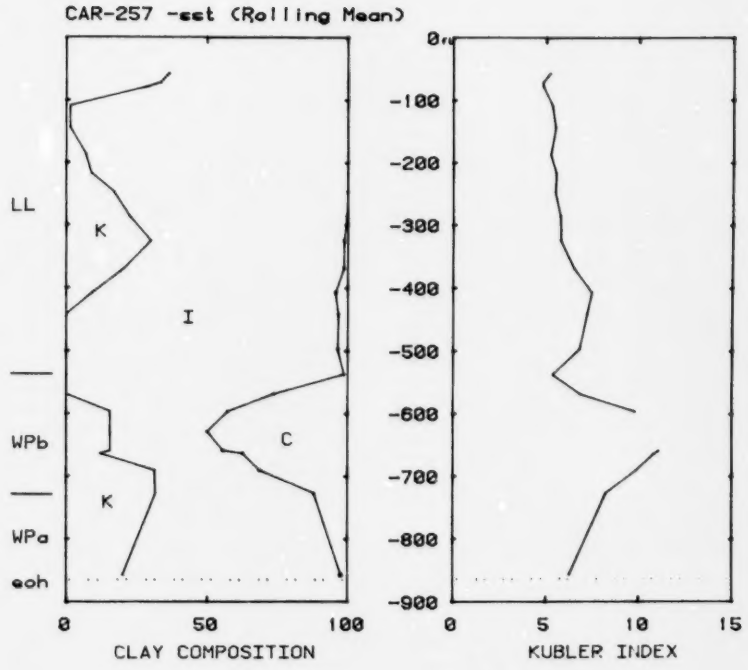

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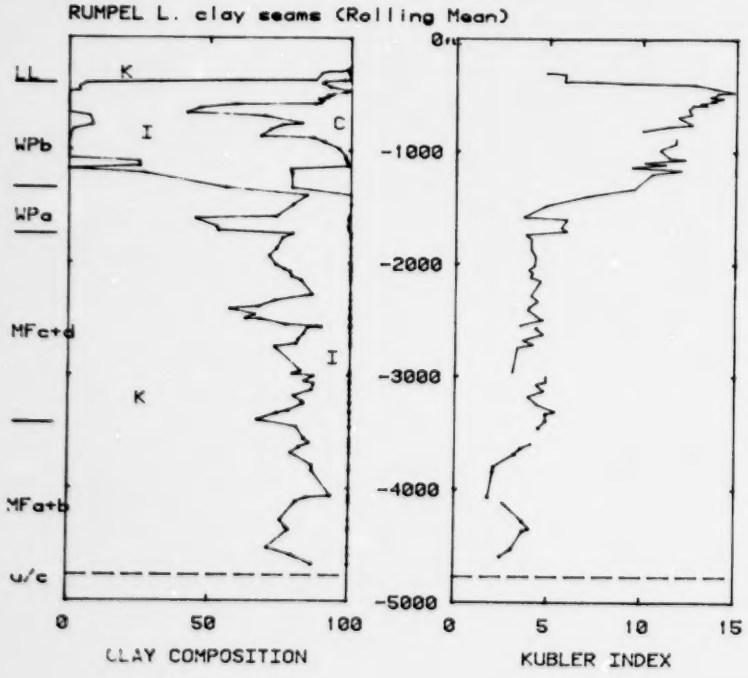

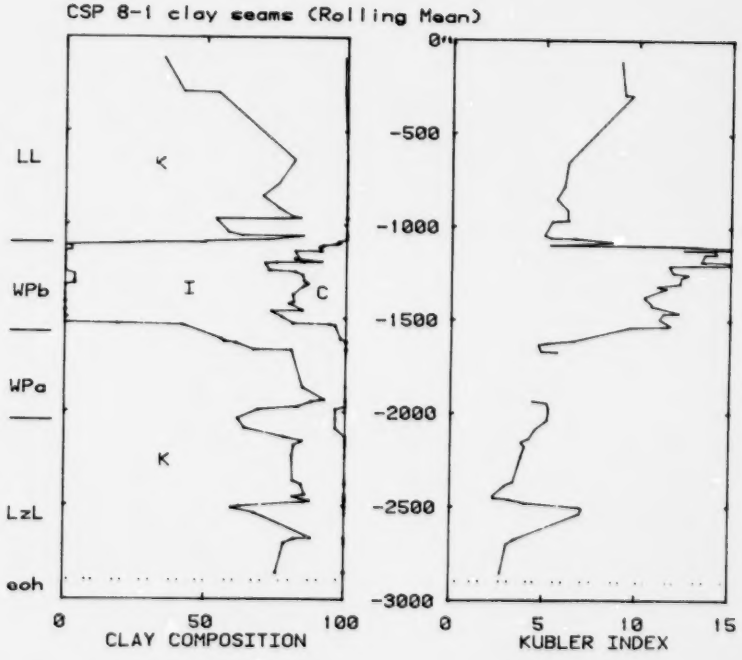

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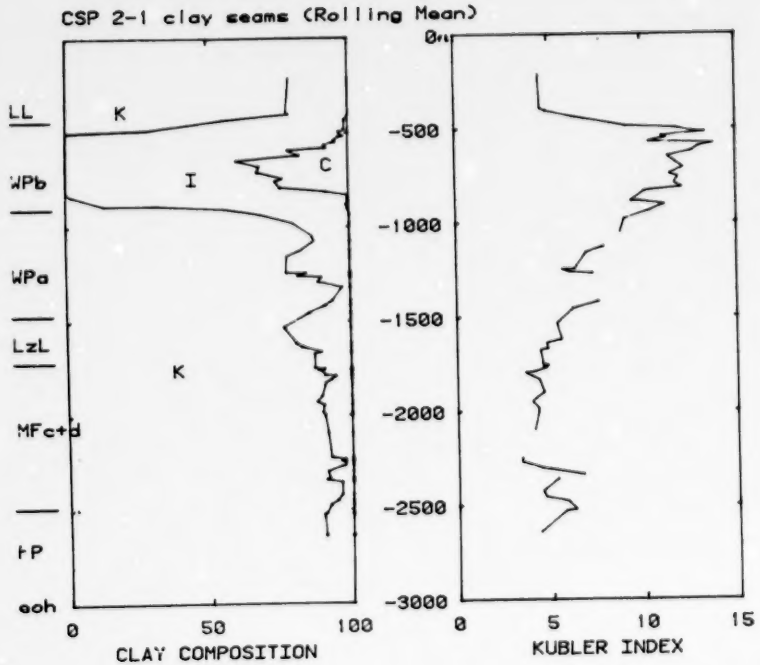

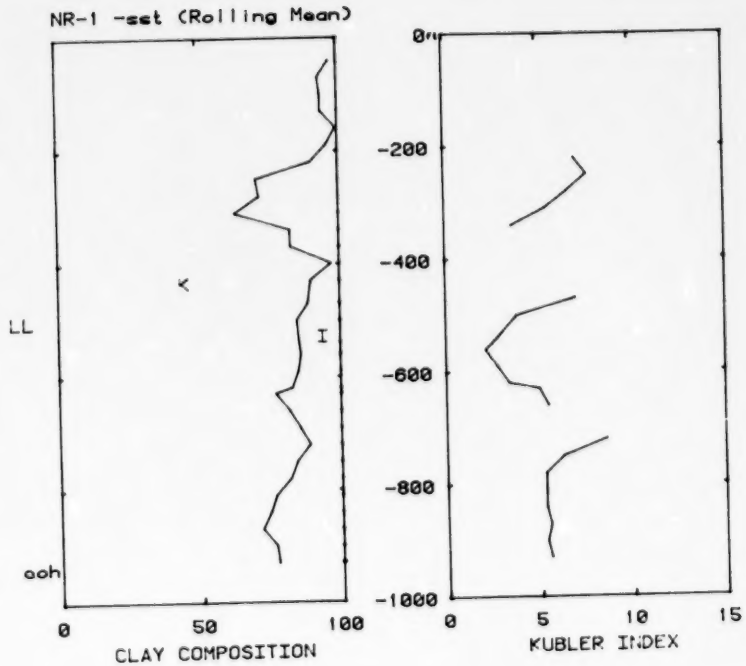

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